Facies analysis and environmental reconstruction of the Upper Cretaceous Chalk Formation in South Limburg, the Netherlands

By

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Abstract

The Late Cretaceous sediments in the Dutch part of the Campine Basin (South-Limburg, the Netherlands) show a great variation in lithology. Since the 1800s, subdivisions of the sediments have been created based on lithostratigraphy, biostratigraphy and chronostratigraphy. In order to understand the depositional environment, these classifications are not sufficient and a facies model is necessary. In this study a facies analysis is presented based on the lithostratigraphic subdivision of Felder (1975) which is matched with an existing biostratigraphic subdivision based on benthic foraminifera. By creating a database for individual members of all outcrops and boreholes in the study area, a dense network of data points can be visualized showing the distribution of lithology per individual member. From this a facies model per individual member is created, showing its distribution and thickness variation, combined with literature data of its age and fossil content. An interpretation of the depositional environment is then created per unit. The facies analysis shows that there is a discrepancy in depocenter for the older Santonian to Campanian sediments compared with the Maastrichtian biodetrital carbonates. It also appeared that the different formations are partly the coeval equivalent of each other. This is the case for the sediments called the ‘Zandig Krijt van Benzenrade’ and the lower part of the Gulpen Formation, as well for the upper part of the Gulpen Formation and lower part of the Maastricht Formation. The facies analysis also gives detailed insights of the tectonics of different structural elements that have influenced these marine shelf deposits of South-Limburg.

Content

1. Introduction

Cretaceous strata are widespread throughout the world. The strata were deposited in a great variance of lithology and different facies can be distinguished. In the south of the Netherlands a succession of Upper Cretaceous sediments is found with a great variation in lithology. This was also the place where Dumont (1849) conducted fieldwork, followed by the introduction of the time stage Maastrichtian. It is also the only area in the Netherlands where Upper Cretaceous deposits are found in outcrops. Since outcrops can be thoroughly studied these can serve as an analogue for Chalk deposits at depth in the subsurface of the Netherlands. In this study a facies classification of the Upper Cretaceous sediments is introduced, based on existing lithostratigraphic and biostratigraphic classifications. The study area comprises the southern part of the Dutch province of Limburg (Fig. 1.1.1), as well as the area in between the city of Visé in the southwest (Belgium) and the city of Aken in the east (Germany).

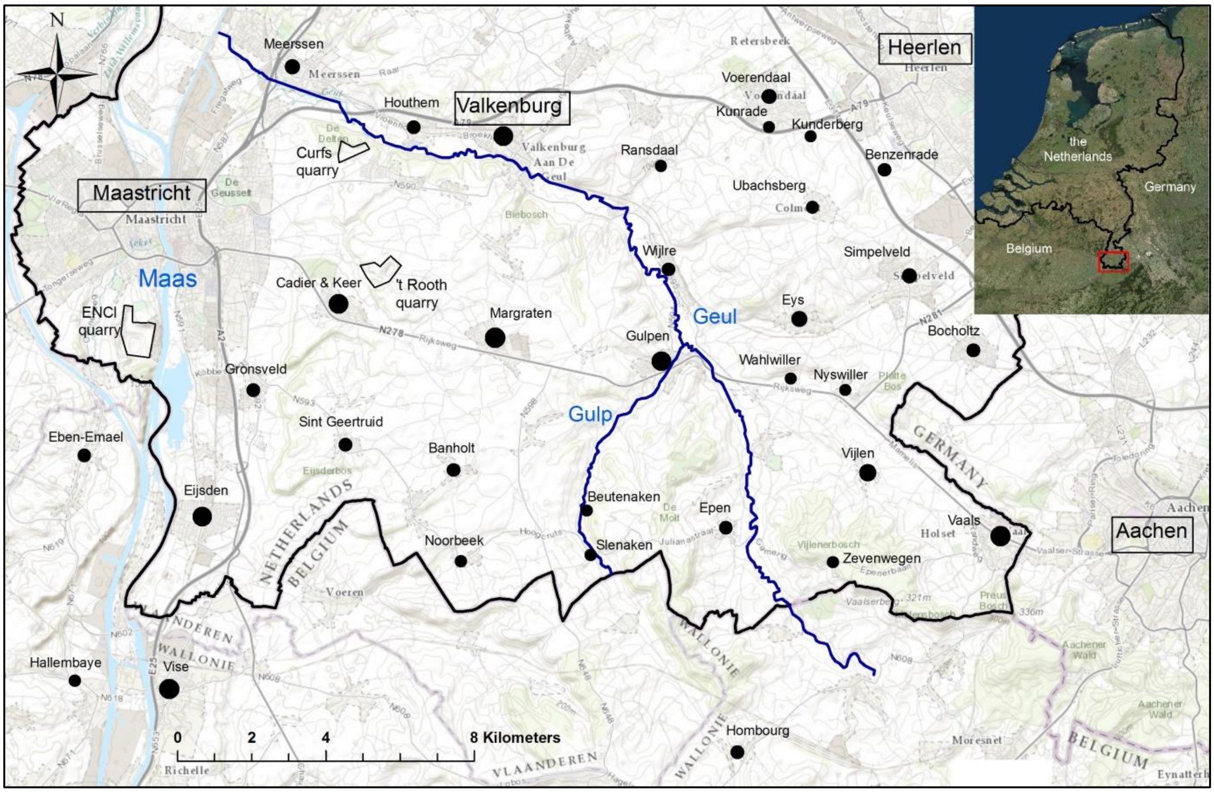

Fig. 1.1.1 Geographical overview of the study area, as well as localities mentioned in this report. (Satellite images and topographic map from ESRI).

The Cretaceous deposits are divided in four different lithostratigraphic formations (Fig. 1.1.2). The Aken Formation is the oldest and its sediments were deposited during the Santonian (Felder, 1975). This is followed by the Vaals Formation which is of Campanian age. During the end of the Campanian the Gulpen Formation formed which continued during the Lower Maastrichtian. The top of the Cretaceous (Upper Maastrichtian) sediments belong to the Maastricht Formation (Felder, 1975). All formations are subdivided in members (Felder, 1975). Within these formations and members the lithology changes laterally and vertically, often very gradual.

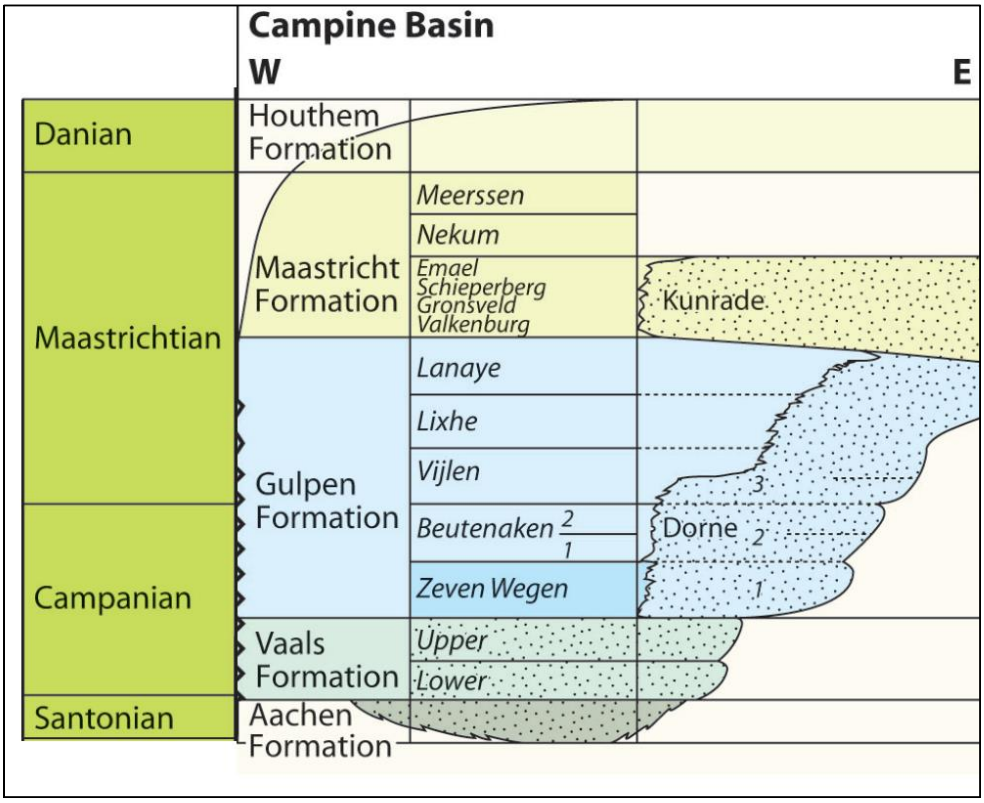

Fig. 1.1.2. Stratigraphic scheme of the sediments in the study area, called the Campine Basin. (modified after Vejbæk et al., 2010)

The classification of these deposits has always been a difficult task. Different classifications have been created throughout the centuries. One can create a lithostratigraphic, a biostratigraphic and a chronostratigraphic subdivision. The lithostratigraphic classification however, does not incorporate time lines and only makes a classification based on lithology. The chronostratigraphic subdivision is only based on timelines. The biostratigraphic subdivision is based on fossil content which also indicates the type of depositional environment of the sediments. By combining all these stratigraphic subdivisions a facies classification (based on lithology and fossil content) can be created. By doing so, facies are described per member(s) of the classification of Felder (1975). This is followed by an interpretation of the depositional environment of the different facies units.

1.1 Previous work

Several authors have tried to make a correlative lithostratigraphic subdivision of the Upper Cretaceous chalk. The English geologist Fitton (1834) was the first to recognize three different units in the Cretaceous strata around Maastricht. Uhlenbroek (1912) created a lithological subdivision that has been used for more than 50 years (Romein, 1962). The problem of this subdivision was that chronostratigraphic terms were applied to purely lithologic units. Schmid (1959) was the first to create a chronostratigraphic subdivision by using fossil rostra of belemnites, and correlating with the Upper Cretaceous belemnite zonation according to Jeletzky (1951). Van der Weijden (1943) and Hofker (1966) published biostratigraphic subdivisions of the strata. Hofker (1966) distinguished several fossil assemblage zones based on benthic foraminifera for the Late Cretaceous (Vaals, Gulpen, Maastricht, Houthem Formations). However, he also recognized rapid changes in foraminiferal assemblages caused by lateral changes in the depositional environment. To solve this problem he introduced a western and an eastern facies for the Maastricht Formation. Felder (1975) introduced a new lithostratigraphic subdivision with type areas and reference sections for the Maastricht area, he distinguished twenty-four lithological units and twenty-nine marker horizons. He also distinguished three different facies in the Maastricht Formation: Maastricht Facies, Schaelsberg Facies and Kunrader Facies. These units pass into each other laterally. Felder (1975) was not the first to recognize this gradual facies change in the Maastricht Formation. Dumont (1849) already regarded the Kunrader limestone as a chronostratigraphic equivalent of the Maastricht limestone.

Robaszynski et al. (1985) created an ecostratigraphic subdivision based on bioclasts and ostracodes that matches the lithostratigraphic and biostratigraphic subdivision of Felder (1975) and Hofker (1966). Felder et al. (1985) suggested that the existing lithostratigraphic subdivision (Felder, 1975) and the foraminiferal zonation (Hofker, 1966) could also be used in northeast Belgium. However, in the eastern Campine area (probably also in the coal-mining area of South-Limburg) the lithology of the Vaals, Gulpen and Maastricht Formations is different than that of the Maastricht area. By means of biostratigraphy he correlated the sediments in the Maastricht area with other locations. Felder et al. (1985) proposed a new term for these sediments that differ in lithology but show resemblance in relative abundance of different fossils. They introduced the term: the Pre-Valkenburg Strata for the sediments in the eastern Campine area. The Pre-Valkenburg Strata are regarded as the equivalent of the base of the Gulpen Formation (Felder et al., 1985).

After the introduction of the term Pre-Valkenburg, Felder et al. (1985) also stated that the upper part of the Gulpen Formation is the lateral equivalent of the lower part of the Maastricht Formation. The same accounts for the upper part of the Vaals Formation that is described as the ‘Zandig Krijt van Benzenrade’. These sediments are presumably the equivalent of the Gulpen Formation (Felder et al., 1985).

1.2 Aim

This paper will combine available studies and datasets in order to come up with a comprehensive description of the Cretaceous strata in South-Limburg and their sedimentological significance.

To do so, a new facies classification will be created. This is based on lithology, fossil content and sedimentary structures. By combining these different elements a facies division is constructed. The facies classification is bounded by horizons of members described by Felder (1975), this is done since this lithostratigraphic classification is matched with biostratigraphic classifications of Hofker (1966) and Robazynski et al. (1985). This suggests that every lithostratigraphic member was spatially deposited in the same time period. However, it appeared that this is not valid for all members. By constructing a facies model the relation between both subdivisions can be better understood. Each facies will be interpreted in terms of depositional environment resulting in a facies analysis. By synthesizing the facies interpretations, a sedimentological history of the study area will be established.

1.3 Geological setting

The study area is located southwest of the Roer Valley Graben, northwest of the Ardennes Massif, and east of the London-Brabant Massif (Fig. 1.3.1). Strictly speaking, the study area is located on top of the London-Brabant Massif, since Cretaceous strata overlie Paleozoic strata.

The Roer Valley Graben formed in the Late Oligocene as a northwest trending branch of the Rhine Graben system (Ziegler, 1988). Before the initiation of the Roer Valley Graben, the northwest - southeast structure was already present, as it has been a fault system since Paleozoic times

(Demyttenaere, 1989). The oldest known sediments in the Roer Valley Graben are of Carboniferous age, unconformably overlain by sediments of Permian, Triassic and Jurassic age. Middle to Late Jurassic and Early Cretaceous sediments are missing, which corresponds to the Late Kimmerian tectonic phase (Geluk et al., 1994). In the Late Cretaceous and Early Paleocene, northwest Europe was subject to two phases of compression (Ziegler, 1990). This is caused by compressional intra-plate stresses exerted by compression in the south of Europe and north Atlantic rifting which resulted in regional inversion of extensional basins (Ziegler, 1990). The Sub-Hercynian inversion phase occurred in the Santonian and Campanian, followed by the Laramide inversion phase in the Maastrichtian to Middle Paleocene (Ziegler, 1990). It is unclear whether the Laramide inversion phase has influenced the Roer Valley Graben.

The London-Brabant Massif is composed of a crystalline basement formed by the Cadomian Pan-African Orogeny (~600 Ma) that resulted in the Avalon Composite Terrane (Guterch et al., 2010). During the Caledonian orogeny (Late Ordovician to Early Silurian ~435 Ma), the Avalon Composite Terrane collided with Baltica as a result of the closure of the Tornquist Ocean. To the south, Caledonian thrust-and-fold structures were produced during the Late Silurian while the new continent Laurussia was formed (Ziegler, 1990; McKerrow et al., 2000). Conversion continued in the Devonian and Carboniferous during the Variscan Orogeny (Cocks et al., 1982). It was in this time period that the Ardennes started to form (Demoulin, 1995). Since the Variscan Orogeny, the London-Brabant Massif formed a topographic high resulting in a period of non-deposition. The oldest sediments that crop out in the study area are of Carboniferous age which were uplifted by the post-uplift of the Ardennes.

Deposition on the south side of the London-Brabant Massif started during the Cenomanian with siliciclastic and carbonatic sediments (Dusar et al., 2007). During the same time, transgressions from the northern basins towards the south were blocked by the inverted Roer Valley Graben and uplifted London-Brabant Massif (Ziegler, 1990). During the Santonian to Campanian, tectonic relaxation of the London-Brabant Massif resulted in a transgression over the crystalline basement linking the Mons Basin in the south with the Campine Basin (Marlière, 1954).

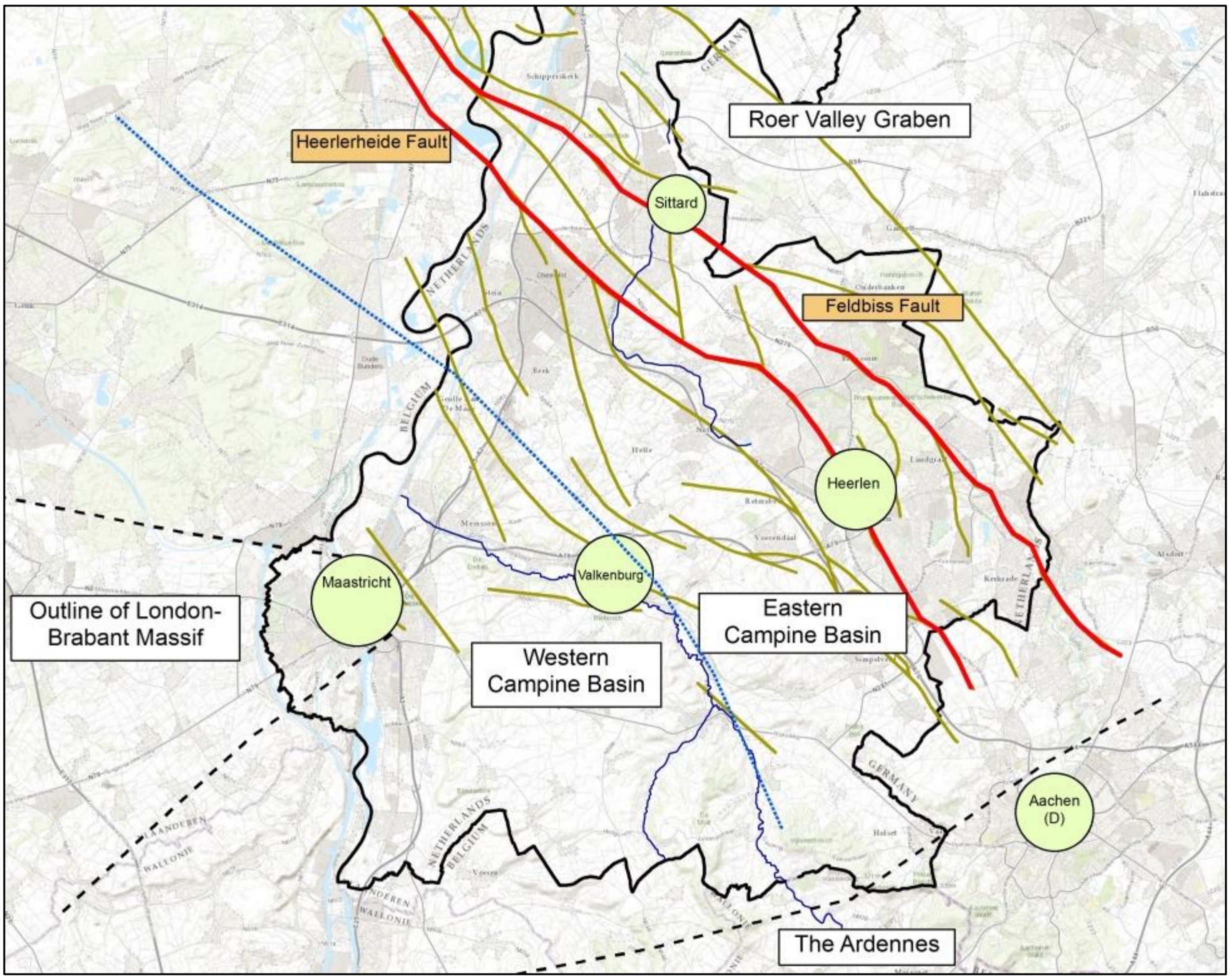

Fig. 1.3.1. Structural elements in the study area. In green circles: cities, in white squares: name of structural elements, in orange squares: names of important faults. (Topographic map from ESRI).

The oldest Cretaceous sediments are of Santonian age and consist of shallow marine siliciclastics and fluvial sediments (Batten et al., 1988). These sediments are grouped as the Aken Formation by Felder (1975). The Aken Formation is covered by shallow marine sediments, deposited in the Campanian (Schmid, 1959 and Jagt et al., 1987). These sediments are named the Vaals Formation (Albers, 1974; Felder, 1975). In the Late Campanian the basal sediments of the Gulpen Formation are deposited as chalk. In the Early Maastrichtian, flint nodules become more common and eventually form flint nodule layers. The Maastricht Formation was deposited in the Late Cretaceous and is characterized by calcarenites with flint nodules in the lower part.

The formations of Felder (1975) can be subdivided in members (fig. 1.3.2). Each member shows a specific horizon that marks the transition to another member. The members and formations of Felder (1975a) are also linked with older subdivisions of Uhlenbroek (1912) and Hofker (1966). Especially the subdivision of Hofker (1966) is interesting since the benthic foraminifera used for the subdivision are also facies dependent. So the combined biostratigraphic subdivision of Hofker (1966) and the lithostratigraphic subdivision of Felder (1975a) are in fact a proto-facies-subdivision.

			Oost van de Maas	West van de Maas	Uhlenbroek (1912)	Hofker (1966)
Formatie v. Houthem		Kalksteen van Geleen	Vc			R
		Kalksteen van Bunde	Vb			Q
		Kalksteen van Geulhem	Va	XIw		P
Formatie v. Maastricht	Boven	Kalksteen van Meerssen	IVf	Xw	Md	N M L
		Kalksteen van Nekum	IVe	IXw	Mc	K
	Onder	Kalksteen van Emael	IVd			I
		Kalksteen van Schiepersberg	IVc	VIIIw	Mb	H
		Kalksteen van Gronsveld	IVb			
		Kalksteen van Valkenburg	IVa			G
Formatie v. Gulpen	Boven	Kalksteen van Lanaye	IIIg	VIIw	Cr4	F
		Kalksteen van Lixhe 3	IIIf	VIw		
		Kalksteen van Lixhe 2	IIIe	Vw	Cr3c	E
		Kalksteen van Lixhe 1	IIId	IVw	Cr3y	
	Onder	Kalksteen van Vylen	IIIc	IIIw		D C
		Kalksteen van Beutenaken	IIIb		Cr3b	B
		Kalksteen van Zeven Wegen	IIIa	IIw	Cr3a	A
Formatie v. Vaals	Boven	Zand van Terstraeten	IIf			
		Zand van Beusdal	IIe			
	Onder	Zand van Vaalsbroek	IId			
		Zand van Grenspaal 7	IIc	Iw	Cr2	A'
		Zand van Cottessen	IIb			
		Zand van Raren	IIa			
Formatie v. Aken		Zand van Aken	Ib		Cr1	
		Klei van Hergenrath	Ia			

Horizons
Horz. van Lutterade
Horz. van Geleen
Horz. van Bunde
Horz. van Vroenhoven
Horz. van Caster
Horz. van Kanne
Horz. van Laumont
Horz. van Lava
Horz. van Romontbos
Horz. van Schiepersberg
Horz. van St. Pieter
Horz. van Lichtenberg
Horz. van Nivelle
Horz. van Boirs
Horz. van Halembaye 2
Horz. van Halembaye 1
Horz. van Wahlwiller
Horz. van Bovenste Bos
Horz. van Slenaken
Horz. van Zeven Wegen
Horz. van Terstraten
Horz. van Beusdal
Horz. van Vaalsbroek
Horz. van Overgeul
Horz. van Grenspaal 7
Horz. van Cottessen
Horz. van Raren
Horz. van Schampelheide
Horz. van Hergenrath

Fig. 1.3.2. Lithostratigraphic subdivision of the Upper Cretaceous and Danian in the study area, as well as the lithostratigraphic subdivision from Uhlenbroek (1912) and the biostratigraphic subdivision of Hofker (1966). (From: Felder, 1975)

2. Methods

The lithostratigraphic subdivision with formations and members (Felder, 1975) has been used to make a facies model for the Gulpen and Maastricht Formations. This is possible since it appeared that every member has a different lithology which is laterally continuous to some degree. The lithostratigraphic subdivision of Felder (1975) has been correlated with a chronostratigraphic subdivision (Schmid, 1959 and Robaszynski et al., 1985) and a biostratigraphic subdivision (Hofker, 1966; Felder P. et al., 1985). This suggests that every member of Felder (1975) contains its own facies and its own age. This is only partly true since many members show a lateral change in lithology and hence different facies can be distinguished. In order to study the change in lithology and to distinguish different facies, information about lithology is needed in many different locations. This data comes from boreholes, as well as outcrops. Additional data comes from literature research.

The Aken, Vaals and Houthem Formation are not subdivided in facies units in this study. The Aken Formation is already subdivided based on lithology and this appears to be a consistent classification in the study area. The Vaals Formation was thoroughly described by Albers (1974) who already recognized two different facies. The Houthem Formation is only exposed in a small area and hence no changes in lithology can be observed. The Houthem Formation is present in many boreholes but it appeared to be impossible to make a distinction solely based on borehole descriptions. The Aken, Vaals and Houthem Formation are briefly presented in this study since the sedimentological history of the study area cannot be explained without.

2.1 Datasets

The most important data consist of boreholes and data from outcrops. The data were obtained from TNO, Geological Survey of the Netherlands (formerly RGD). The RGD (‘’Rijks Geologische Dienst’’) archived all reports, borehole descriptions and investigations of the Dutch subsurface. In the study area a local office of the RGD was located: the Geologisch Bureau Heerlen (GB). Most data of the study area were stored in the archives of the Geologisch Bureau Heerlen. Heerlen received its own office due to the presence of the Carboniferous mines that eventually closed in the 1960s and 1970s. The Geologisch Bureau Heerlen closed in 1997 after the Rijks Geologische Dienst merged with TNO.

Boreholes
All boreholes that reach the Cretaceous strata have been studied and re-interpreted. Many have only been interpreted on formation scale and not on member scale. The reinterpretation resulted in additional data of the distribution and thickness of individual members. The quality of borehole descriptions greatly varies and not all are useful for this study. The oldest boreholes are from the 1900s and the quality of their descriptions is often inferior. The descriptions of the lithology are derived from cuttings in the borehole that were transported to the surface. A problem with this method is the possibility of contamination with sediments from younger stratigraphic layers. So one should keep in mind that exact boundaries are difficult to pinpoint.

Most valuable for this study are boreholes from the 1960s till the 1980s. These have been described in detail and supply detailed information about lithology on decimetre to metre scale. Many of these boreholes have also been examined on foraminifera, and allocated to one of the foraminifera zones of Hofker (1966). These zones have been linked to the subdivision of Felder (1975). Since most zones have specific guide benthic foraminifera species, these can also be used to make a facies classification. Benthic foraminifera are more sensitive to changes in depositional environment than, for example, planktic foraminifera and hence more useful to reconstruct the depositional environment (Hofker, 1966).

Outcrops
The number of outcrops in the study area has decreased considerably during the last decades. The late W.M. Felder has visited most of the outcrops in Cretaceous deposits of South-Limburg. Felder was an employee of the RGD and worked at the GB in Heerlen, together with his colleagues he made an inventory of all outcrops in South-Limburg and neighbouring locations in Belgium and Germany. Outcrops were logged and an identification number was assigned, together with the coordinates of the outcrop. The stratigraphy of the deposits was interpreted and they were subdivided into members. Before 1975 the lithostratigraphic subdivision of Uhlenbroek (1912) was used, and since 1975 the

lithostratigraphic subdivision of Felder (1975) became the standard. Many locations have also been documented in sketches which emphasize lithological characteristics. Some outcrops have also been sampled and interpreted based on benthic foraminifera using the classification of Hofker (1966).

In some cases Cretaceous sediments crop out in quarries, which have been extensively studied in the 1980s and 1990s by the RGD. The descriptions of these quarries were often accompanied by measurements of $CaCO_3$ content since the quality of the mined carbonates is highly dependent on their $CaCO_3$ content. Where possible, these measurements have been included in this study since they show the relative amount of siliciclastic grains which is an indication of the proximity of a source area. Erosion only occurred where sediments were reworked by waves or where sediments were exposed to aerial processes, in other words: nearshore or land.

Additional fieldwork has been conducted in the winter of 2015-2016. Outcrops were visited with the aim to observe sedimentary structures. After some visits to outcrops it appeared to be impossible to observe sedimentary structures without damaging the outcrop as a fresh surface was needed. Since many outcrops are on private property or geological monuments it was not allowed to damage them. However, the presence of many subsurface limestone mines offered an opportunity to observe sedimentary structures without damaging them, these observations have been included in this study.

Literature
The sediments in the study area are very prone to weathering. The only opportunity to describe and log outcrops is in quarries that are actively mined. Unfortunately mining has come almost completely to an end. This means that some observations of specific members must come from scientific literature and reports. Data from reports of the RGD and GB have been used for additional data of sediments.

Profiles and descriptions of Felder et al. (1981, 1985, 2001), who created a new classification of ecozones by grouping different bioclasts in the sediments, are also used. The term bioclasts is used for carbonitic remains of organisms with a size of 1 to 2.4 mm. It was discovered that the composition of bioclasts greatly varied vertically in a sequence, but remained constant over large distances in a lateral sence (Felder et al., 1985). His division in ecozones is based on relative abundance, and not absolute abundance, of a specific group of bioclasts. The main bioclast groups are: 1. Foraminifera, 2. Bryozoans, sponges and corals, 3. Molluscs and brachiopods, 4. Echinoderms, and 5. Rest group. The relative abundance of each of these groups is used to distinguish different facies in the same member. This is possible since Felder et al. (1981, 1985 and 2001) logged many different boreholes throughout the study area, by using these profiles and logs the fossil content can be used to interpret the depositional environment of the distinguished facies.

2.2 Synthesis of data

The datasets were combined and a database was created in which every individual member in a specific location is described by using borehole or outcrop descriptions. In total 1265 data points from outcrops and 2052 data points from boreholes were retrieved. Each data point represents an individual member in a specific location. One specific location may contain multiple data points as it is common that multiple members are reached or exposed. Table 2.2.1 shows the elements of each data point.

Element of data point	Description
ID	Identification number of outcrop or borehole
X-coordinates	X-coordinates according to Rijksdriehoekscoodinaten (RD)
Y-coordinates	Y-coordinates according to Rijksdriehoekscoodinaten (RD)
Member or Formation	Corresponding member or formation (Felder, 1975)
Top layer NAP	Top of member relative to NAP
Base layer NAP	Base of member relative to NAP
True thickness	Thickness of member when bounded by Cretaceous strata above and below
Min. thickness	Minimum thickness of member when base is not reached or covered by Cenozoic strata (Oligocene)
Description member	Lithologic description and additional information from borehole report or fieldwork at outcrops (for example: $CaCO_3$ content, sedimentary structures and Hofker (1966) zones)

Table 2.2.1. Elements of each data point.

All data points are spatially visualized in ArcMap (GIS software). By doing so, distribution and thickness maps were created per member of formation. The thickness measured is the true vertical thickness, this however, is not the same as the true stratigraphic thickness of a member. But due to an inclination of only 1˚ towards the northwest, the true vertical thickness roughly equals the true stratigraphic thickness.

The distribution and thickness maps are used to describe the different facies distinguished in this report. Since every data point contains a description of the lithology, the distribution of changes in lithology can be visualized per member or facies unit. Descriptions of sedimentary structures from outcrops are also included, as well as additional information ($CaCO_3$ content and benthic foraminifera zones of Hofker, 1966). This is supported by two cross-sections (Appendix) created from boreholes and outcrops from the southwest to the northeast (Roer Valley Graben). Together with distribution and thickness maps the cross-sections visualizes trends per member. After the description and distribution of the facies unit, information about fossil content is presented that is used to interpret the depositional environment. Information about fossil content comes from literature, mainly from Hofker (1966) and Felder (1981, 1985 and 2001) and Felder et al. (2000). The description, distribution and fossil content of the facies unit are then used to interpret the depositional environment. This is accompanied by interpretations by other authors.

3. General characteristics of the Upper Cretaceous deposits

Felder (1975a) defined horizons that bound the members in the Gulpen and Maastricht Formations. Two different transitions are found: 1. a hardground or cemented layer covered by a basal fossil-grit layer and 2. a distinct flint-nodule layer. Since all members contain one of these transitions, its genesis is explained in this chapter to avoid repetition for each individual unit.

3.1 Diagenetic features

Hardground

Throughout the Cretaceous strata hardgrounds occur regularly, often developing to maturity towards the margins of basins (Håkansson et al., 1974). A hardground forms when sediments are cemented by calcite. They may form horizontal horizons that can be tracked for kilometres (Felder et al., 2000). Hardgrounds mark submarine lithification events that are associated with hiatuses (Scholle et al., 1983). Lithification occurred due to the precipitation of carbonate cement as a result of bacterial metabolism in anoxic redox zones below the paleo sediment surface, up to several decimetres deep (Bromley, 1975; Bischoff and Sayles, 1972; Zijlstra, 1995). The degree of lithification varies in the sediments. The most complete lithification resulted in the formation of a hardground, which is mineralized by cement, encrusted and bored (Voight, 1974). Zijlstra (1995) stated that most strongly lithified layers formed when the deposition rate was zero and the sediment resided in the anoxic redox zones for a long time. This is the case when newly deposited sediments are continuously removed. In other words, the hydrodynamic energy of the water was high enough to erode and transport sediments on the sea floor. Since not all intervals are equally cemented, an external forcing is likely to have operated on the sediment surface with varying magnitude. Zijlstra (1995) attributed this to the increase of hydrodynamic energy conditions during storms. He recognized a cyclicity of 20,000 years (precession) in which the average hydrodynamic energy conditions during storms changed from a minimum to a maximum. During an increase of storm intensity and related hydrodynamic energy, the increased erosion rates equalled the subsidence rates and the deposition rate was consequently zero resulting in a lithified layer (Zijlstra, 1995). During an increase in storm intensity, the sediment that had been lithified below the sediment surface, was repeatedly eroded, exhumed and exposed during storms (Zijlstra, 1995). During maximum storm intensity, the lithified sediment was continuously exposed and a a bored and encrusted hardground formed.

Flint

Flint forms out of a chemical process that dissolves silica from detrital skeletal opal and precipitates as early diagenetic silica polymorphs in or just outside the most anoxic redox zones of sulphate and carbondioxide reduction (Bromley, 1975 and Zijlstra, 1989). These zones of precipitation occur around burrows and just below and parallel to the paleo sea floor (Bischoff and Sayles, 1972; Zijlstra, 1995). When the precipitation of early diagenetic silica continued, a semi planar layer formed. This however, can only occur when a specific layer remains in the zone of anoxic redox reduction. So in order to continue the precipitation of silica the deposition rate of new sediments should be zero or very small. As seen in the formation of hardgrounds, this condition is met when one assumes a strong variation in hydrodynamic energy conditions during a 20 ky storm cyclicity (Zijlstra, 1995). During these conditions anoxic redox zones occurred in the same layer for a long period of time since no new sediment was deposited due to the high hydrodynamic energy conditions during storms. Consequently, specific layers have increased silica concentration while other layers lack the zones of concentrated silica precipitates due to continuous sedimentation.

Late diagenetic precipitation of flint takes place where there already is an increased concentration of early diagenetic silica polymorphs (Zijlstra, 1995). During this process, individual flint nodules keep growing and eventually form a nodular flint layer. These nodular flint layers have formed during a period when sedimentation rate was low or zero and hence the early diagenetic silica concentration was inversely proportional to the deposition rate (Zijlstra, 1995).

3.2 Sedimentary structures

Fossil-grit layer
A coarse fossil-grit layer contains coarse broken fragments of bioclastic sand. The bioclastic sand is composed of skeletal remains of fossils. In many locations it is also characterized by siliciclastic grains (quartz), Carboniferous clasts, intraformational clasts (up to tens of centimetres in diametre), glauconite, phosphate and pyrite.

The broken bioclastic grains indicate deposition and possibly transport during high energy conditions. This is shown by the coarse grain size suggesting deposition during relatively high energy conditions. Terrigenous siliciclastic grains indicate a more proximal terrestrial environment as terrigenous clasts must have been transported from the shoreline. Intraformational clasts are also common, and point to erosion of indurated carbonate layers that are redeposited. Again, this is an indication of high energy conditions which suggests that nearby carbonates were cemented and eroded. These clasts appear to be of local origin and originate from cemented layers (hardgrounds) nearby. During storms these cemented layers may have been eroded and redeposited in a matrix of bioclastic sand.

According to Zijlstra (1995) the glauconite, phosphate and pyrite have replaced carbonate clasts. These minerals were also observed in intergranular pore space where they precipitated, which suggests these are authigenic minerals. Bromley et al. (1975) and Zijlstra (1995) suggested that these minerals precipitated in redox zones due to bacterial metabolism and dissolution of detrital minerals. This is the same chemical process as for carbonate and silica precipitation (anoxic redox conditions), but occurs in the suboxic zones of manganese and iron reduction. A high concentration of glauconite and pyrite at the base that is gradually decreasing upwards indicates that the minerals are presumably first eroded and then redeposited. Zijlstra (1995) also recognized this and suggested a sorting in a fining upwards sequence according to their fall velocity. An increase in authigenic mineral concentration could suggest an increased rate of glauconisation due to reworking (Burst, 1958). Another possibility is that these minerals were eroded in other locations, transported and redeposited.

Cross bedding
Cross bedding is far more abundant in the Aken and Vaals Formation than in the chalk of the Gulpen and Maastricht Formation. Cross bedding is commonly found at the base of cycles or members (Felder, 1975; Zijlstra, 1995). The general lack of visible cross bedding is caused by the thorough bioturbation during deposition (Zijlstra, 1995). When deposition occurred faster than burrowers could rework the sediments, sedimentary structures were preserved. These conditions were met during waning of storms when tempestites were deposited (Zijlstra, 1995). Structures as wavy lamination or cross bedding are found in the Maastricht Formation, and are described by Zijlstra (1995).

3.3 Cyclicity

Most members classified by Felder (1975) show a basal fossil-grit layer ranging from a few centimetres to decimetres. Below, a hardground is commonly found indicating a period of non-deposition before deposition of the basal fossil-grit layer. The top surface of the hardground is often undulating (Fig. 3.3.1). The fossil-grit layer is always accompanied by an increase in size of the bioclasts. Above the fossil-grit layer, grain size gradually decreases.

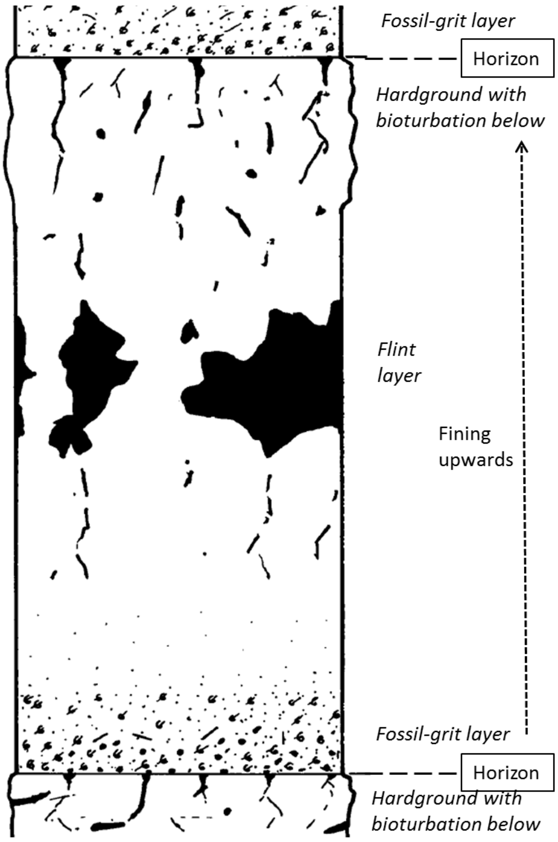

Fig. 3.3.1. Sketch of an ideal cycle. The base consists of a few decimetre thick fossil-grit layer. In the middle part a flint layers has formed in and around burrows. In the top a hardground has formed, below, burrows are still visible. The cycle ends with an erosional surface followed by a new fossil-grit layer. Vertical scale ranges from a few decimetres to multiple metres. (Sketch modified after Felder, 1976)

The observed outcrops often do not show this ideal case. Often the hardground is not well developed or partly eroded, flint is not present everywhere and the fossil-grit layer did not always form.

4. Facies model

First the Aken and Vaals Formation will be summarized, the distribution and thickness of these formations will be presented. The depositional environment is then interpreted based on lithology and sedimentary structures. This is followed by a facies analysis of the different members of the Gulpen and Maastricht Formations of which the depositional environment is interpreted. The Cenozoic Houthem Formation will again only be summarized.

4.1 Aken Formation

The Aken Formation consists of the oldest Mesozoic sediments in the study area, deposited in the Santonian (Batten et al., 1988). Felder (1975) subdivided this formation in three members that are shortly described upwards:

- The Member Klei van Hergenrath represents the oldest sediments which are characterized by grey silty clays, occasionally alternated with coarse-grained sands. Organic remains are common, for example remains of trunks, plants and leaves. The sediments also contain many pollen and seeds of terrestrial flora (Albers et al., 1979). Seatearths are also common in clay layers, as well as organic rich layers that formed into lignite (Albers et al., 1979).

- The Member Zand van Aken is characterized by fine-grained and well-sorted sand with locally thin clay lenses (Albers et al., 1979). Some sand layers are cemented and form hard continuous (hundreds of metres) sandstone beds. A cyclicity can be observed with channel infills, cross bedding (Fig. 4.1.1 and 4.1.2) and intraformational clay nodules at the base and lamination towards the top (TNO - RGD, report no. 21a, 1971). Cross bedding shows that flow directions have changed on a regular basis (TNO - RGD, report no. 21a, 1971). Quantitative research of the different flow directions could provide detailed information about the depositional environment. The lower part of the member is characterized by cycles with large channel infills of multiple metres wide and decimetre to metre thick cross bedding, bioturbation is absent (TNO - RGD, report no. 21a, 1971). Towards the top of this member channels and cross bedding decrease in size and intervals with bioturbation are common. Intervals of thin clay layers in the cross bedding (Fig. 4.1.2) become more common towards the top (TNO - RGD, report no. 21a, 1971). Organic remains of vegetation are still present but less common as in the previous described member (Felder et al., 2000).

Fig. 4.1.1. Channel infills, cross bedding and lamination in the Member Zand van Aken in quarry Käskorb in Belgium (ID: 62D0074). Vertical scale: 5 to 10 m. (Photo from RGD, 1994)

Fig. 4.1.2. Cross bedding and cementation in the Member Zand van Aken in quarry Käskorb in Belgium (ID: 62D0074). Note the thin intervals of clay layers in the fine-grained sand dominated cross bedding. Vertical scale: several metres. (Photo from RGD, 1994)

- The Member Zand van Hauset is characterized by fine-sand, in the top locally alternated with clay layers (Albers et al., 1979). Felder et al. (2000) recognized a cyclicity in these sediments with a fine-grained sand with high clay content at the base, followed by fine-grained sand with local cemented sandstone nodules, and bioturbation in the top. Cross-bedding is only occasionally present while lamination and bioturbation is common (Fig. 4.1.3). Large channels of widths of multiple metres can be distinguished (Fig. 4.1.4). This member is only present in the southeast of the study area.

Fig. 4.1.3. Bioturbation destroyed wavy lamination in the Member Zand van Hauset in quarry Bingeberg in Belgium (ID: G62G0069). Vertical scale: 1 to 1.5 m. (Photo from RGD, 1991)

Fig. 4.1.4. Large channel infill, cut on top by laminated beds. Member Zand van Hauset in quarry Bingeberg in Belgium (ID: G62G0065). Vertical scale: 10 to 12 m. (Photo from RGD, 1991)

The distribution of the Aken Formation is bounded by the Heerlerheide fault in the north and east (Fig. 1.1.1 and 4.1.5). Carboniferous strata are reached by boreholes but no Aken Formation was encountered north of this fault. The thickness of the Aken Formation is largest in the east, near the Heerlerheide fault (Fig. 4.1.6). Profile 1 (Appendix) clearly shows a gradual increase in thickness of the Aken Formation towards the northeast. The thickness ranges from less than 1 m to 78 m. In the east, the thickness exceeds the average thickness found in the rest of the study area that is commonly only a few metres to 15 m thick.

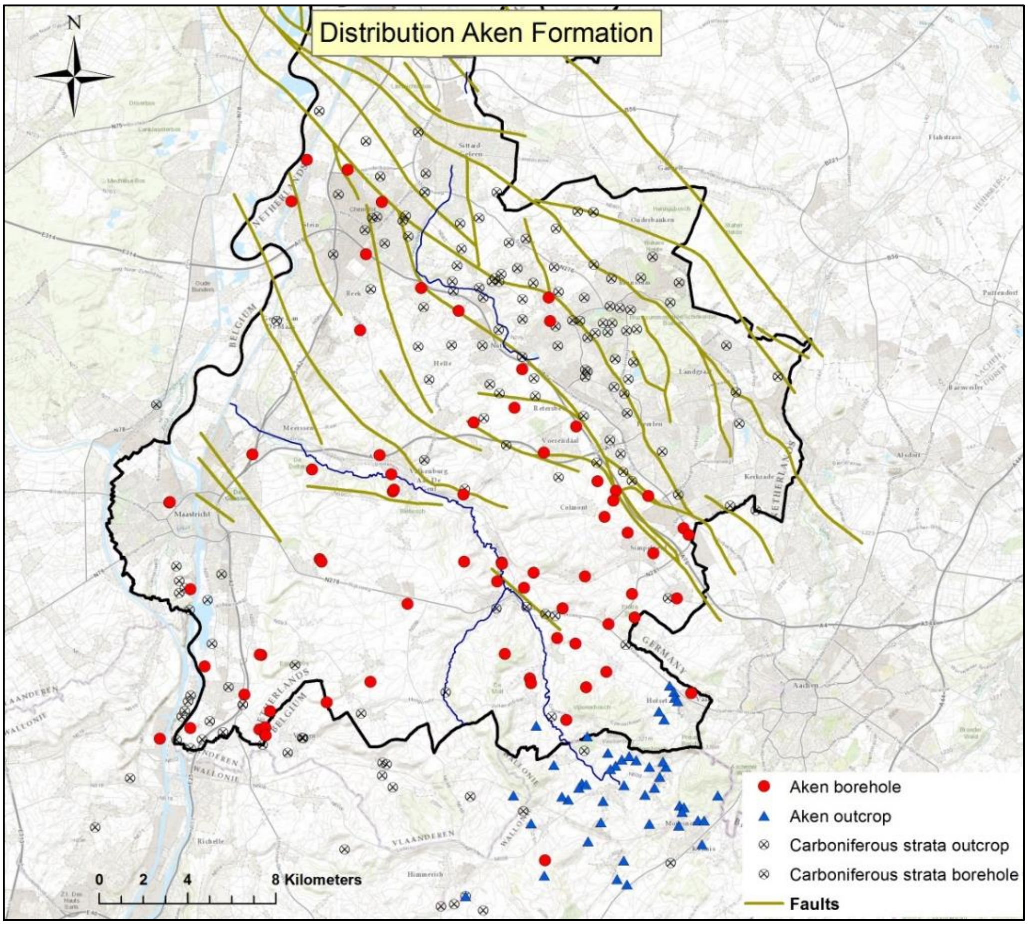

Fig. 4.1.5. Distribution of locations where the Aken Formation is present.

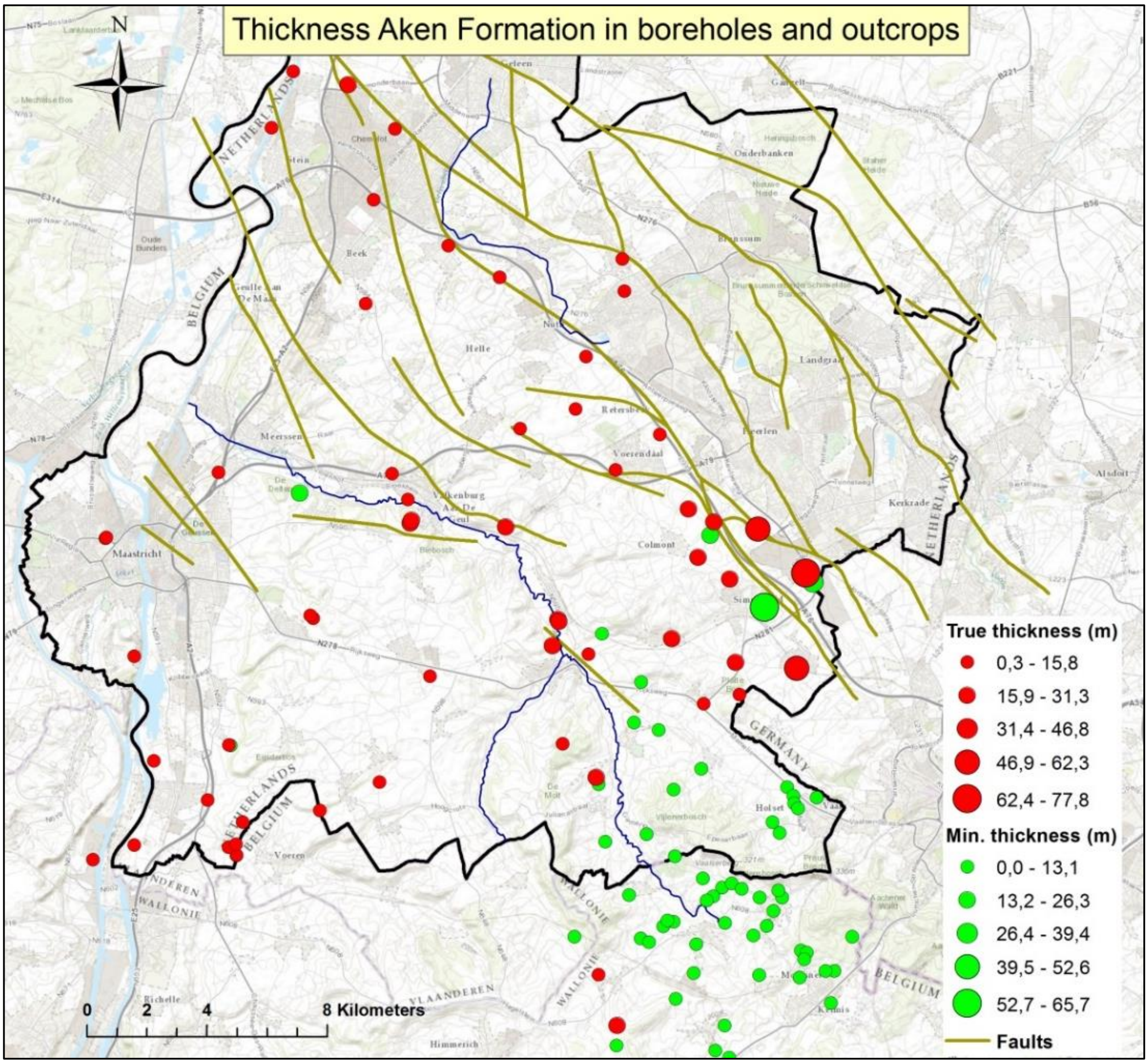

Fig. 4.1.6. Thickness of the Aken Formation. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Interpretation of depositional environment

Every member was deposited in another depositional environment since the lithology and sedimentary structures vary.

The Member Klei van Hergenrath can be interpreted as a dominantly fluvial depositional environment with floodplains (organic-rich clay layers, seatearths, peat layers), locally alternated with small channels (very coarse-grained sands). This is supported by the high abundance of pollen and seeds that indicate terrestrial conditions or at least nearshore conditions. Bless et al. (1987) interpreted these clay and silt deposits alternated with lenses of lignite as lagoonal swamp deposits.

The Member Zand van Aken can be interpreted as a dominantly marine depositional environment with high hydrodynamic energy conditions and above the fair-weather wave base as the abundance in cross-bedding shows. Bless et al. (1987) interpreted the cross-bedded sands with well-rounded quartz grains and locally bioturbation as beach deposits (Bless et al., 1987). Towards the top, cross bedding sets are alternated with thin clay layers which can be interpreted as tidal bundles which shows that the depositional environment was subject to tidal influence. Felder et al. (2000) interpreted the paleo-water depth during deposition around 20 m. The presence of (drifted) plant remains, pollen and seeds suggests that the shoreline was nearby.

The Member Zand van Hauset can be interpreted as a marine depositional environment with episodic influence of waves and alternating limnetic and brackish conditions (Albers et al., 1979). Hydrodynamic energy conditions are decreasing and periodically non-deposition conditions prevail resulting in cementation and bioturbation. Felder et al. (2000) interpreted the paleo-water depth as 20 to 40 m, based on the lack of tidal influence and episodic presence of wave influence. This member is only present in the southeast of the study area (southeast of the line Epen – Vaals). In other locations it is probably not recognized since this member can only be distinguished in outcrops. In boreholes the lithology is the same as the underlying member and a distinction is difficult to make. Since the uplift of the Ardennes is strongest in the southeast, these sediments are only exposed in the southeast.

The gradual change from terrestrial to shallow marine is caused by a marine transgression throughout the deposition of the Aken Formation. Second- or third-order sea level changes may cause the cyclicity in the sediments that represent small regressions and transgressions of the marine environment (Haq et al., 1988). The largest thickness of the Aken Formation is found in the east, south of the Heerlenheide Fault. North of this fault, the Aken Formation is absent. This suggests that the Roer Valley Graben had the largest accommodation space during deposition of the Aken Formation. At the end of the Santonian compressive forces resulted in inversion of the Roer Valley Graben (Bless et al., 1987). It appears that the Heerlerheide Fault was subject to a large amount of offset since the Aken Formation is absent towards the north while it was deposited in a rift basin (Roer Valley Graben). The large accommodation space in a rift basin and the largest thickness close to the Heerlerheide Fault suggest that the Aken Formation is most probably eroded during the inversion.

4.2. Vaals Formation

The Vaals Formation consists of marine sediments from the Campanian (Schmid, 1959 and Jagt et al., 1987). Albers (1974) separated this formation is seven members, of which Felder (1975a) adopted six. The sediments consist of calcareous fine-grained sands and silts with varying clay content, as well as varying amounts of authigenic glauconite (Bless et al., 1987). The base of the Vaals Formation always contains a coarse-grained conglomeratic layer with clastic pebbles (Fig. 4.2.1). Albers et al. (1979) recognized a cyclicity in deposits in the east of the study area. The base contains laminated fine-grained sands with silt and clay, locally with fossil-grit layers, followed by fine-grained glauconitic sand with bioturbation in the top (Fig. 4.2.2). Locally cemented sandstone nodules can be found. In the lower part of the Vaals Formation the base of cycles occasionally shows several metres wide channel infills, which are absent in the upper part. The cyclicity fades away towards the west since silt and clay become more dominant and the cyclicity is less well observed (Albers et al., 1979). Towards the top of the Vaals Formation the cyclicity is better developed as burrows are more pronounced at the top of cycles (TNO - RGD, report no. 21a, 1971). In some locations the lower part of the Vaals Formation shows traces of a paleosol, which indicates aerial exposure (Fig. 4.2.3).

Fig. 4.2.1. Boundary between Aken Formation (white color) and Vaals Formation (reddish color). Note the coarse-grained gravel at the base. Photo from outcrop Kasteel Vaalserbroek (ID: 62D0445). Vertical scale: 2 to 3 m. (Photo from RGD, 1982)

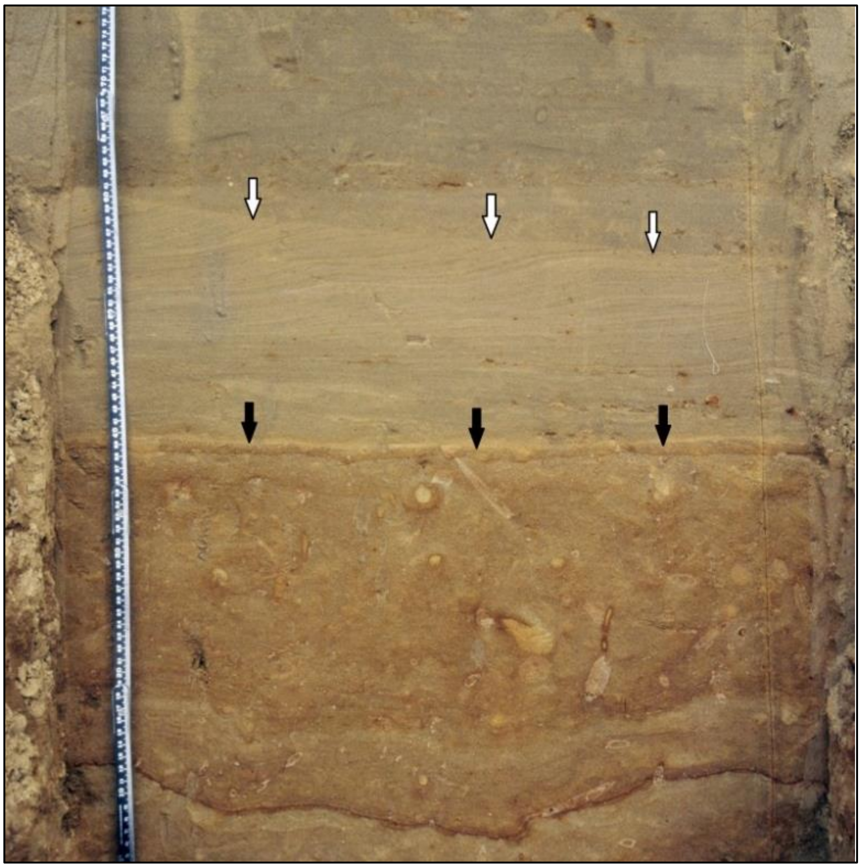

Fig. 4.2.2. Top and base of a sedimentary cycle in Vaals Formation. Underneath layer with black arrows many marine burrows can be observed. Above, centimetre scale lamination that is cut by another feature, transition marked with white arrows. Photo from outcrop Randweg Vaals (ID: 62D0096). Vertical scale: 0.9 m. (Photo from RGD, 1987)

Fig. 4.2.3. Paleosol covered by laminated fine-grained sand of the Vaals Formation. Photo from outcrop Randweg Vaals (ID: 62D0096). Pen for scale. (Photo from RGD, 1987)

Albers (1974) recognized that the members in the marine Vaals Formation show a lateral change in lithology. In the east (south of Vaals, and near Aken) these sediments are mainly composed of fine-grained sand, but towards the west the lithology gradually passes into silt with a large clay content (Fig. 4.2.4). The $CaCO_3$ content varies from 2 to 30% in the study area, with the highest values in the west and northwest of the study area (Felder et al., 2000). Hofker (1966) recognized characteristic benthic foraminifera in the sediments of the Vaals Formation. The specific species were grouped as Zone A', subdivided in Lower, Middle and Upper (Hofker, 1966). This indicates that the sediments were formed during similar environmental conditions.

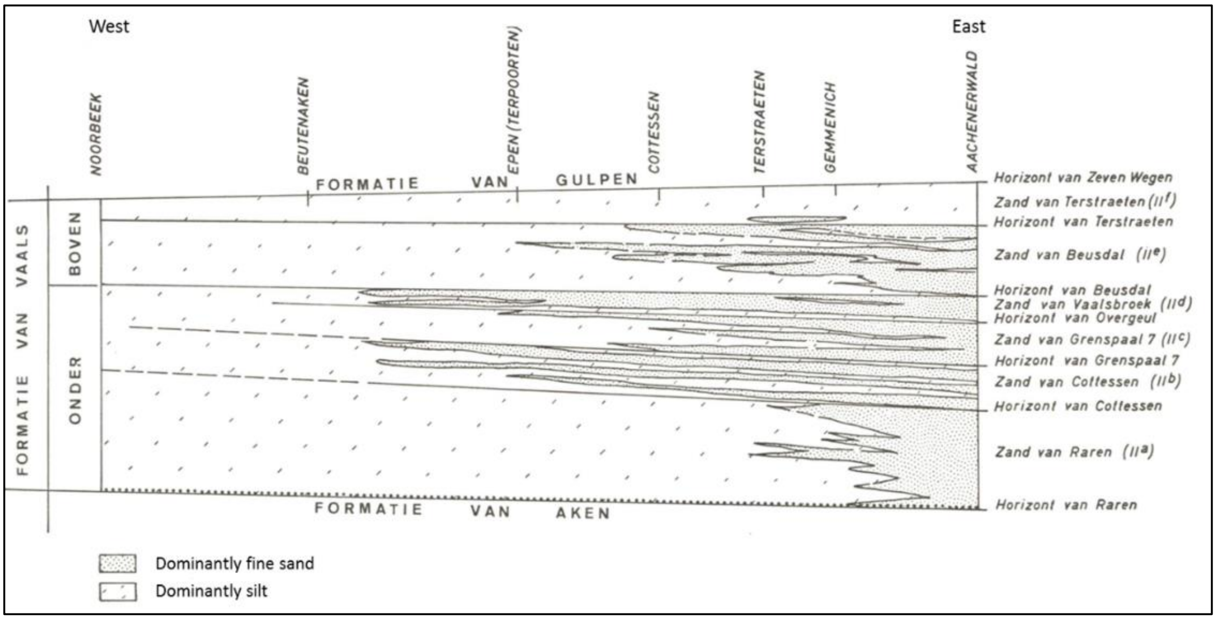

Fig. 4.2.4. Lateral and vertical distribution of the members of the Vaals Formation and their lithofacies. (Figure modified after Albers, 1974)

The distribution of the Vaals Formation is limited by the Heerlerheide Fault in the north and in the east (Fig. 4.2.5). Its thickness is difficult to retrieve since it is not possible to distinguish between the sandy Vaals Formation and the sandy base of the Gulpen Formation found near the Roer Valley Graben.

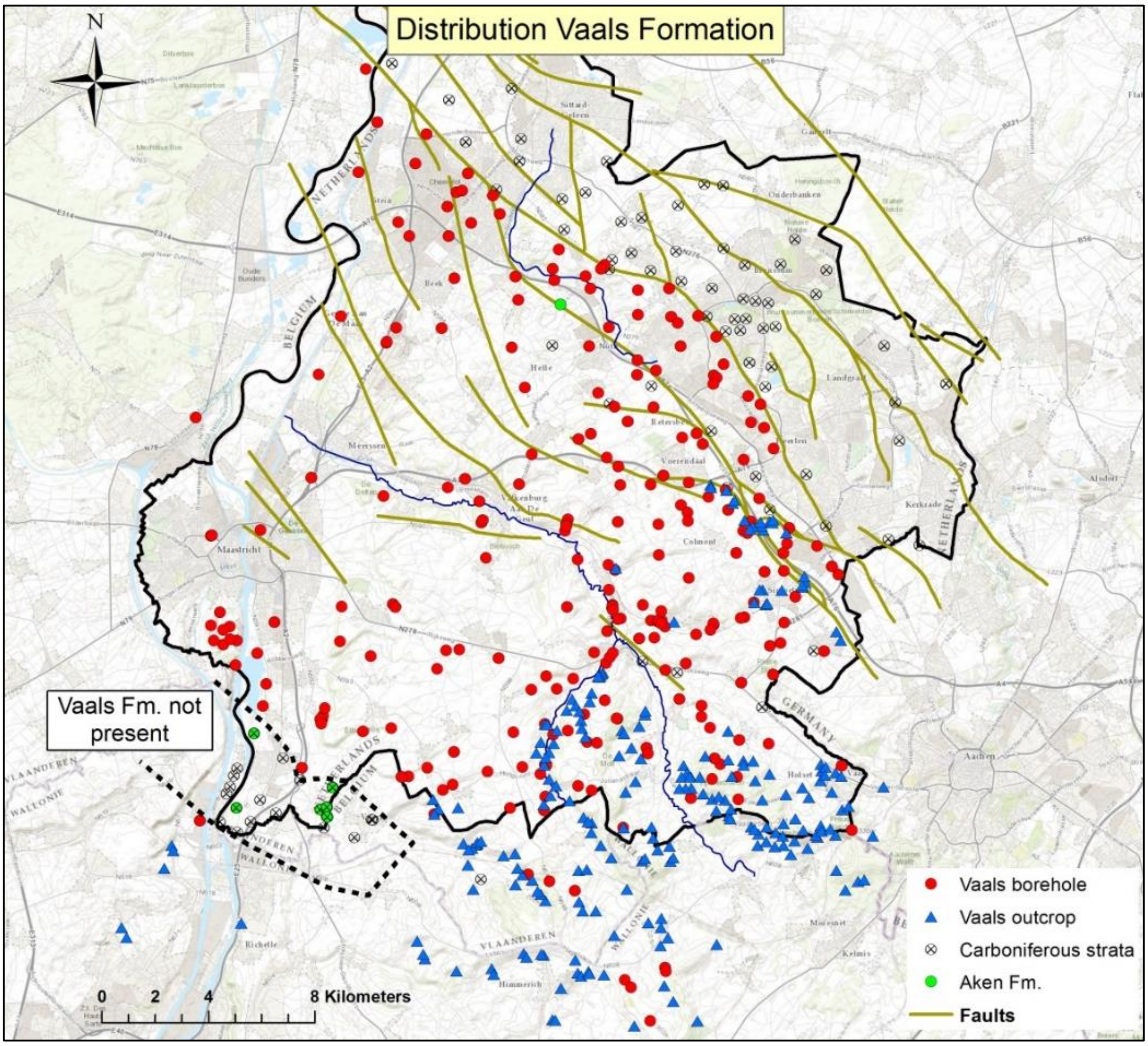

Fig. 4.2.5. Distribution of locations where the Vaals Formation is found.

Interpretation of depositional environment

The lower part of the Vaals Formation is interpreted as a marine depositional environment above the fair-weather wave base (Albers et al., 1979), evidenced by the presence of laminated sediments and infill of channels. Towards the top of the Vaals Formation, hydrodynamic energy conditions decrease, as is shown by the absence of channels, a better preservation of cycles and a gradual decrease in grain size (Felder et al., 2000). The upper part of the Vaals Formation is interpreted to be deposited in an environment with a paleo-water depth of 20 to 50 m, based on the absence of sedimentary structures that indicate flowing water on a regular basis. Albers et al. (1979) interpreted these sediments as deposited below the fair-weather wave base. The abundance of authigenic glauconite and the common occurrence of burrows points to a shallow marine environment where glauconite precipitated around burrows that was later reworked (Zijlstra, 1995).

The gradual change from fine-grained sand in the east to silt and clay in the west suggests that the depositional environment was deepening towards the west, as the hydrodynamic energy conditions were decreasing. Different cycles of transgression and regression of the shoreline can be distinguished within the Vaals Formation (Albers, 1974). Overall, a general transgressive trend is observed. The alternation in fine-grained sand with silt and clay can be interpreted as second- or third-order cycles in sea level change. Laterally, the coarse-grained nearshore facies thin out towards the west where a fine-grained distal facies is more dominant.

4.3 Zevenwegen Facies

The sediments of the Zevenwegen Member of the Gulpen Formation (Felder, 1975) are considered to belong to one facies, which is named the Zevenwegen Facies in this study. The facies is characterized by white chalk with a bioclast grain size of 37 µm (Bless et al., 1987). The basal few metres, contains glauconite of which the concentration gradually decreases upwards (Albers et al., 1979). The top of this member is cemented during diagenesis and forms a hardground with burrows below (Fig. 4.3.1). These burrows are typically filled with green to white coloured carbonates from overlying sediments.

Fig. 4.3.1. The hardground in the top of the Zevenwegen Member. Above the hardground, sediments of the Vijlen Member. Photo from 1986 in quarry CBL, Halembaye, Belgium (RGD).

The distribution of the Zevenwegen Member is limited by the northern line Maastricht – Margraten – Zevenwegen (Fig. 4.3.2). Further north, the Zevenwegen Member has not been observed in outcrops and boreholes. Southeast of Maastricht, the Zevenwegen Member is absent in a several kilometre wide and twenty kilometre long zone that stretches from the NW to the SE (Fig. 4.3.2 and profile 1 in Appendix). The thickness of the Zevenwegen Member ranges between 0 to 33.5 m (Fig. 4.3.3). The greatest thicknesses are found in the southwest of the study area and decreases towards the northeast and eventually thin out.

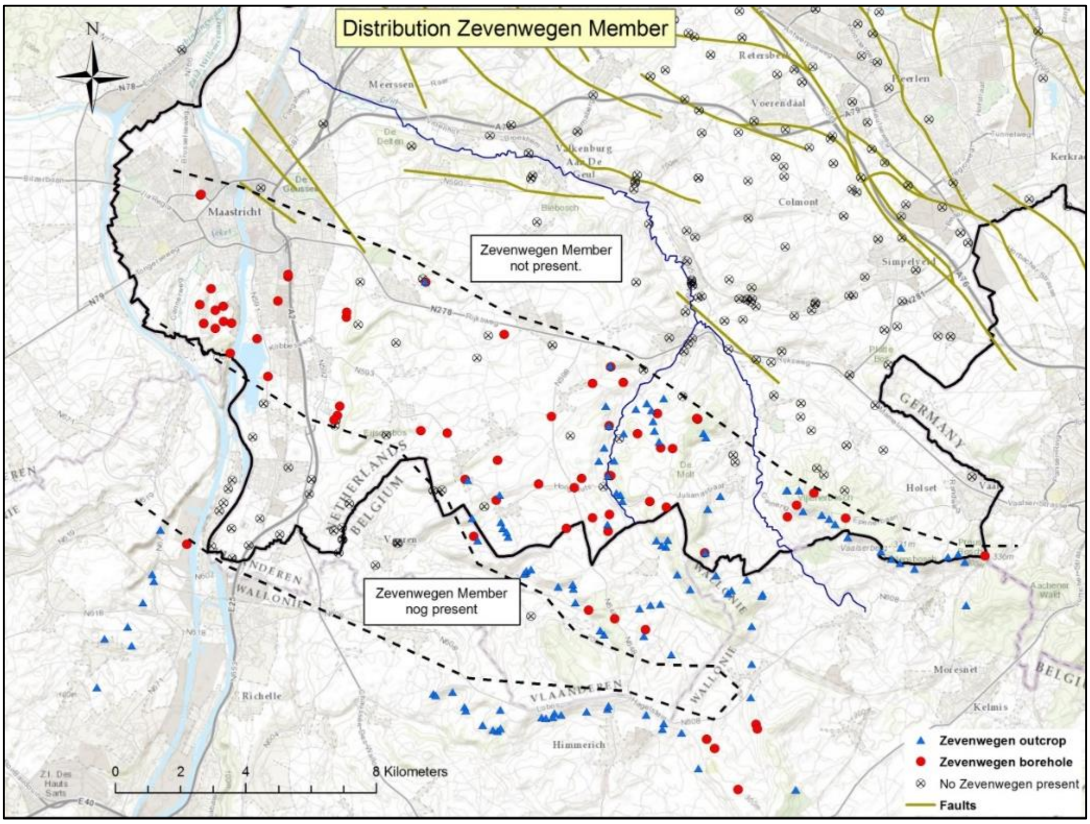

Fig. 4.3.2 Distribution of locations where the Zevenwegen Member is found.

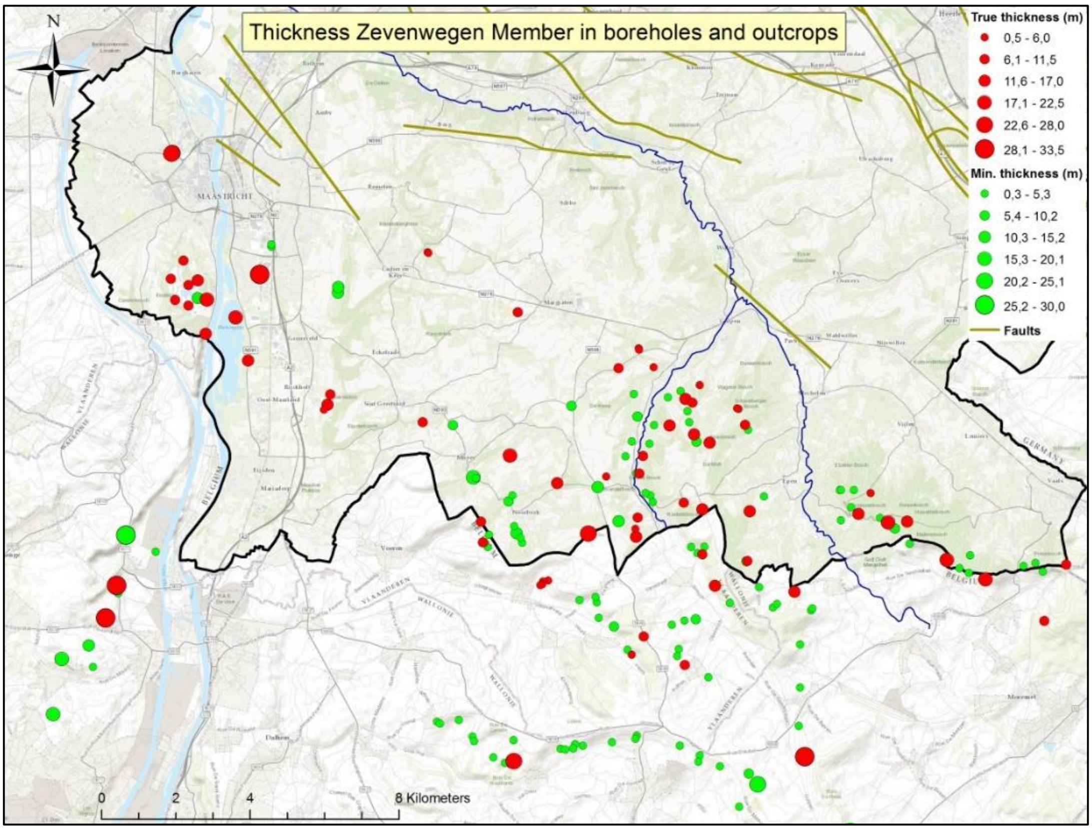

Fig. 4.3.3. Thickness of the Zevenwegen Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations

where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

The average $CaCO_3$ content is highest in the southwest of the study area (Hallembaye), where it reaches up to 97% in the top (RGD, outcrop data 61H019). Towards the north (ENCI quarry), the $CaCO_3$ content gradually decreases to 75% at the base and 85-90% in the top (RGD, Rapport no. OP 5730). According to Felder et al. (2000) the average $CaCO_3$ content is 75 to 80% in the southeast (Vaals and Epen).

Age

The age has been established by the occurrence of specific belemnite species (Schmid, 1959). The sediments are placed in the Late Campanian (Van der Tuuk, 1980; Jagt et al., 1987; Keutgen et al., 1990).

Fossil content

The sediments of the Zevenwegen Member are characterized by specific benthic foraminifera species, grouped as Zone A by Hofker (1966). This indicates that the sediments were formed during similar environmental conditions.

Felder et al. (1985) found a relatively high number of echinoderm spines and remnants of the prismatic layer of bivalves as well as many belemnite guard fragments in the Zevenwegen Member. Albers et al. (1979) stated that the diversity of the microfossils is a factor two higher in Halembaye (southwest of the study area) than in the localities of Zevenwegen and Beutenaken (southeast of the study area). The basal glauconitic layer in a quarry in Hallembaye is enriched in fossils: irregular echinoids, belemnites, sponges, corals and brachiopods (Felder, 1983). Compared to the older Aken and Vaals Formations, the number of ornamented ostracods decreased (Felder et al, 1985).

Felder (2001) shows that the relative abundance of different fossil groups is roughly constant over different locations. There is however a small decrease in the number of bivalves from the southeast (Hombourg) to the southwest (Hallembaye) of the study area. While further west, in Landen (Belgium), the number of bivalves and prismatic bivalves shows an increase. This decrease is also observed from the southwest (Hombourg) to the northwest (Maastricht). Throughout the Zevenwegen Member multiple sudden increases in belemnite numbers can be observed in all locations. The number of belemnites relative to the total fossil content strongly varies, but in all locations a peak in relative abundance can be observed (profiles Felder, 2001).

Interpretation of depositional environment

The sediments of the Zevenwegen Facies were deposited in a relative low energy depositional environment as is shown by the absence of sedimentary structures and the relative small bioclast grain size. This is confirmed by the fossil content. The presence of irregular echinoids in the southwest is an indication for a soft sediment surface, while regular echinoids are generally found on rocky substrates (Brusca et al., 1990). The decreasing number of ornamented ostracods relative to the older Aken and Vaals Formation, is indicative for a decrease in hydrodynamic energy conditions (Felder et al., 1985).

The Zevenwegen Facies is not found in a NW to SE trending zone, younger as well as older sediments are missing in this area (Appendix, profile 1) which indicates that an erosional phase has probably removed these sediments.

The high carbonate content in the southwest (up to 97%) points to a distal depositional environment, as relative few siliciclastic grains were found in the sediments. Towards the north and the east the

$CaCO_3$ content decreases and the sediments contain a higher percentage of siliciclastic grains which points to a more proximal environment.

Towards the northeast of the study area these sediments thin out and a coeval lateral equivalent called the Zandig Krijt van Benzenrade Facies is found (Appendix, profile 1).

4.4 'Zandig Krijt van Benzenrade' Facies

The sediments called the 'Zandig Krijt van Benzenrade' are classified as a specific facies named: Zandig Krijt van Benzenrade Facies. The sediments consists of a fine to medium coarse-grained sand in the lower part, and a fine- to coarse-grained glauconitic sand with hard nodules and lenses of limestone in the upper part. Felder et al. (2000) recognized a cyclicity of grainsize in the sediments. The base of each cycle contains thin glauconitic sand layers that are partly cemented, followed by glauconitic silt with local cementation in the middle and upper part, and a well-cemented top that occasionally forms a hardground (Felder et al., 2000). According to Felder et al. (1989) isolated gravel beds are occasionally present. In some locations, these sediments are described as sandy to silty glauconitic marls with limestone alternations (Bless et al., 1987). Traditionally these deposits were classified as Vaals Formation (Albers, 1974; Felder, 1975), but since the 1980s it is regarded as the equivalent of the lower part of the Gulpen Formation based on biostratigraphy (Jagt et al., 1987; Felder et al., 1989).

It is difficult to recognize this facies unit in boreholes since the sediments show close resemblance with the Vaals Formation. So in borehole descriptions these sediments have been interpreted as Vaals Formation and sometimes as Gulpen Formation. In southwest direction these sediments have been interpreted mostly as Gulpen Formation, as the limestone lenses and beds were better developed. In northeast direction the sediments dominantly developed as fine-grained sand and it is recognized as the Vaals Formation (Fig. 4.1.1). Because of this, only a distribution map is created as a thickness map is not reliable. The cross-section (Appendix, profile 1) shows that the thickness clearly increases towards the northeast. This facies is found north of the line Wahlwiller – Gulpen –Wijlre up to the Benzenrade fault. This facies is also found at the base of the Zevenwegen Member in boreholes in the south of the study area. Its thickness there is only a few metres and gradually passes into the Zevenwegen Facies.

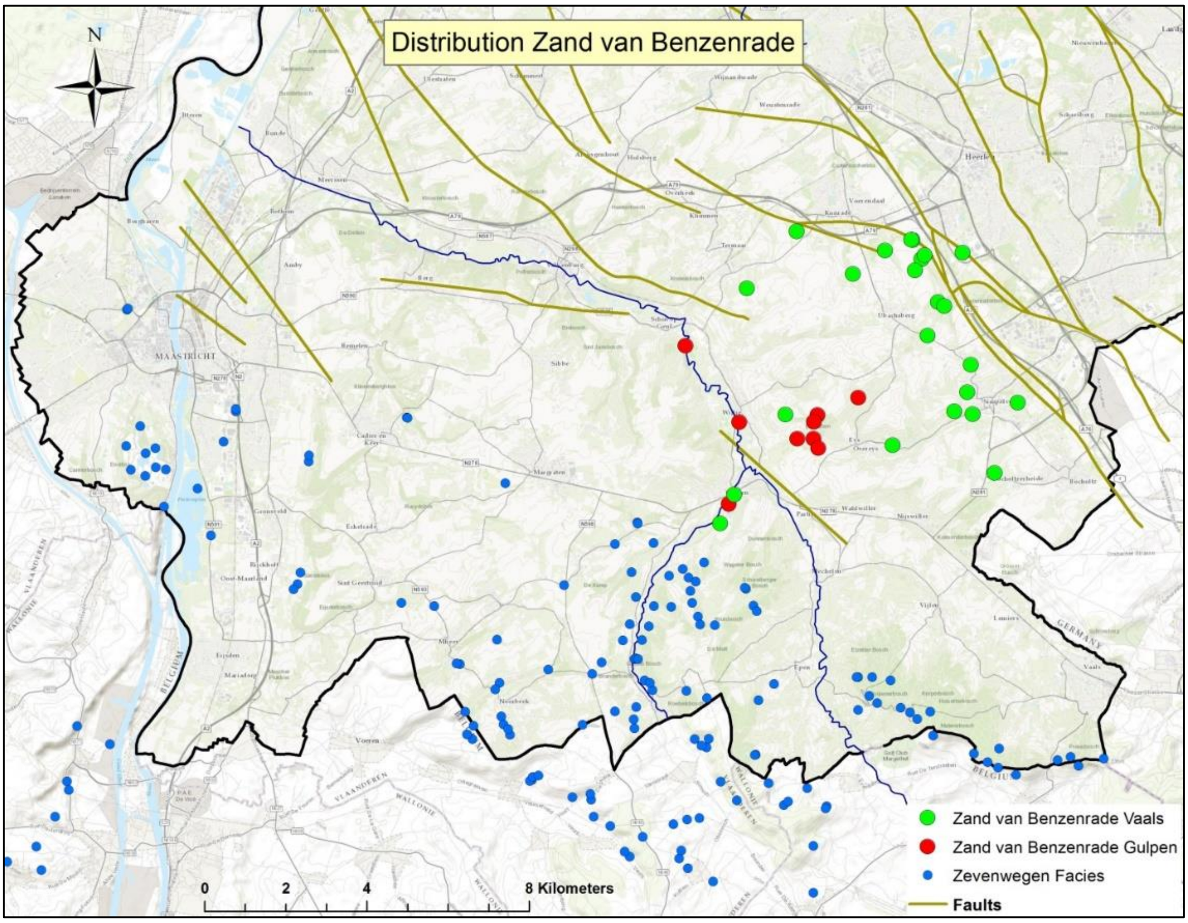

Fig. 4.4.1. Distribution of locations where the Facies Zandig Krijt van Benzenrade is found, as well as its equivalent: the Zevenwegen Facies. This facies is interpreted as the Vaals or Gulpen Formation by TNO and RGD, a reinterpretation of both formation resulted in this distribution map.

Age

The sediments are of Late Campanian age based on belemnites and foraminera zonations (Jagt et al., 1987; Felder et al., 1989).

Fossil content

The sediments of the Facies Zandig Krijt van Benzenrade have always been regarded as the top of the Vaals Formation (Albers, 1974; Albers et al., 1979). Hofker (1966) distinguished specific benthic foraminifera, grouped as Zone A' Upper for these sedimets. Hofker (1966) recognized many species that were characteristic for the Zevenwegen Member, as well as some other species that were absent in the Zevenwegen Member. From this he concluded that these sediments were deposited in another depositional environment. Biostratigraphic research shows that the Zevenwegen Facies and the Facies Zandig Krijt van Benzenrade are coeval (Jagt et al., 1987; Felder et al., 1989; Felder, 2001).

According to Felder et al (1989) the amount of bioclasts increases upwards. Molluscs are most dominant in the bioclast content, as well as smaller amounts of echinoderm (crinoids) clasts and remains of solitary corals in the upper part. Ostracods show relatively few ornamented specimens, compared to older stratigraphic units. Jagt et al. (1987) recognized the presence of large thin-shelled ammonites for the first time.

Interpretation of depositional environment

The large siliciclastic content of the sediments suggest a proximal depositional environment, close to an area where non-carbonate sediments were eroded. Compared to the Aken and Vaals Formation these sediments contain relatively few ornamented specimens, which indicates lower energetic conditions during deposition (Felder et al., 1989). The decrease in energy conditions can be interpreted as a deepening of the depositional environment. This is confirmed by the presence of large thin-shelled ammonites, which is interpreted as a deepening of the depositional environment relative to the older Vaals and Aken Formations (Jagt et al., 1987). This however, is contradictory with the interpretation that the sediments were deposited in a proximal setting. An explanation might be that increased erosion rates due to increased uplift resulted in a large supply of siliciclastic sediments blocking the carbonate factory. The timing of the increased siliciclastic supply is in agreement with the Sub-Hercynian inversion period of the Roer Valley Graben (Ziegler, 1990).

The distribution of the Facies Zandig Krijt van Benzenrade is limited to the northeast of the study area. The thickness (Appendix, profile 1 and 2) increases towards the northeast indicating increased accommodation space towards the Roer Valley Graben. This is contradictory with the tectonic setting as it is expected that the Roer Valley Graben was inverted. Possibly the inversion was limited to the central part of the Roer Valley Graben (north of the Heerlerheide Fault), supplying siliciclastic sediments to the basin. Further south, the uplift was less and most fault blocks could still accommodate sediments.

The sediments of the Zandig Krijt van Benzenrade are not only found as the coeval equivalent of the Zevenwegen facies in the southwest of the study area. Several boreholes show that this facies unit is present below the typical chalk of the Zevenwegen Facies, as is the case in the south of the study area near the Belgian Ardennes. The lithology is identical to the Vaals Formation but the fossil content is different and closely resembles that of the Zevenwegen Facies (Felder et al., 1985; Jagt, 1988). This shows that both facies pass into each other, both laterally and vertically.

4.5 Beutenaken Facies

The Beutenaken Facies corresponds to the Beutenaken Member of Felder (1975).The sediments consist of light grey chalk with a bioclast grainsize of 51 µm (Bless et al., 1987). The basal few metres contain glauconite, which gradually decreases in concentration upwards. The high glauconite concentration causes a slight greenish colour at the base of this member. Small amounts of fine-grained glauconite remain present in these sediments, as well as small flint nodules in the upper part. In the top of Beutenaken Member a hardground formed due to diagenesis (Felder, 1975). In the ENCI quarry (south of Maastricht), the carbonate content varies between 65 and 80% with an average of 71% (RGD, Rapport no. OP 5730).

The distribution of the Beutenaken Member is only limited, it is only found near the ENCI quarry and around Beutenaken and Slenaken (Fig. 4.5.1). Its thickness ranges from 0 to 16.5 m (Fig. 4.5.2). In the Belgian part of the Campine Basin the Beutenaken Member is thicker and more abundant (Felder et al., 1985).

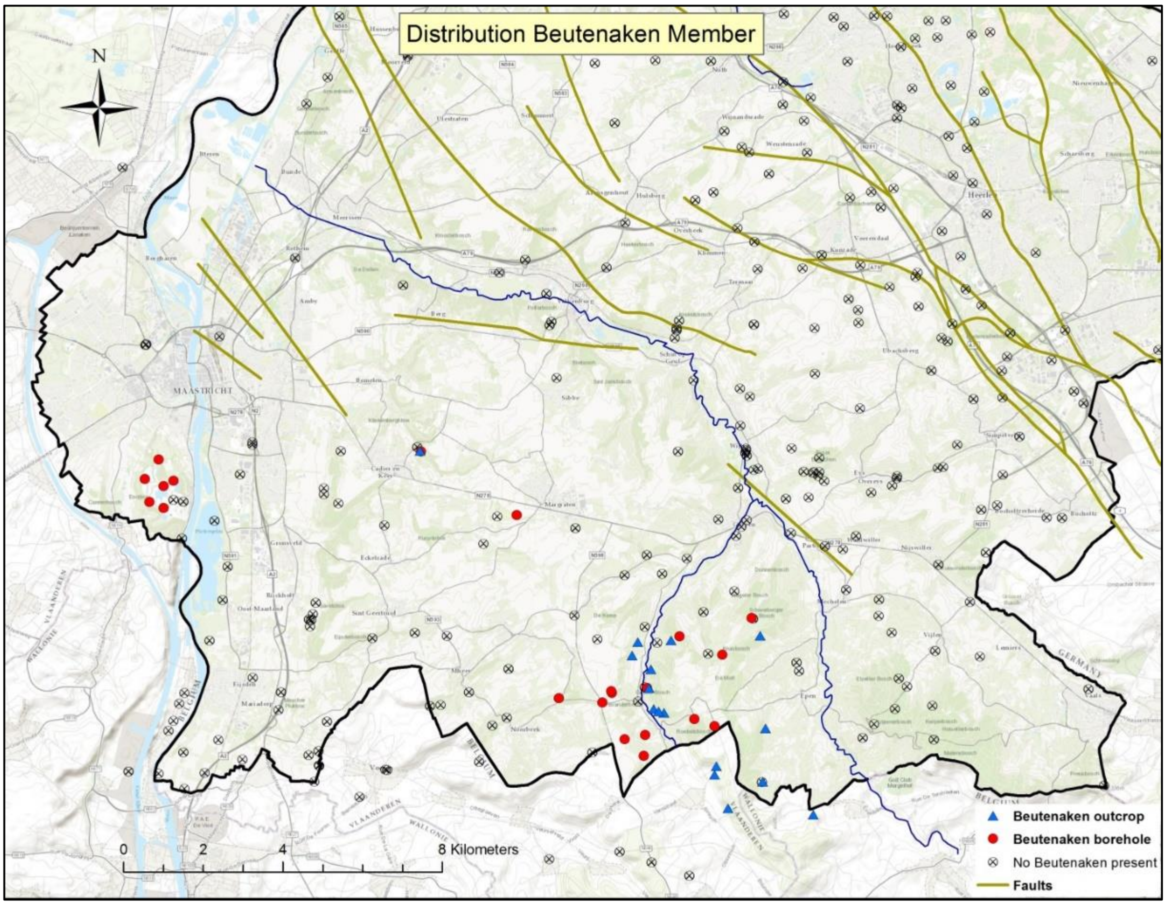

Fig. 4.5.1. Distribution of locations where the Beutenaken Member is found.

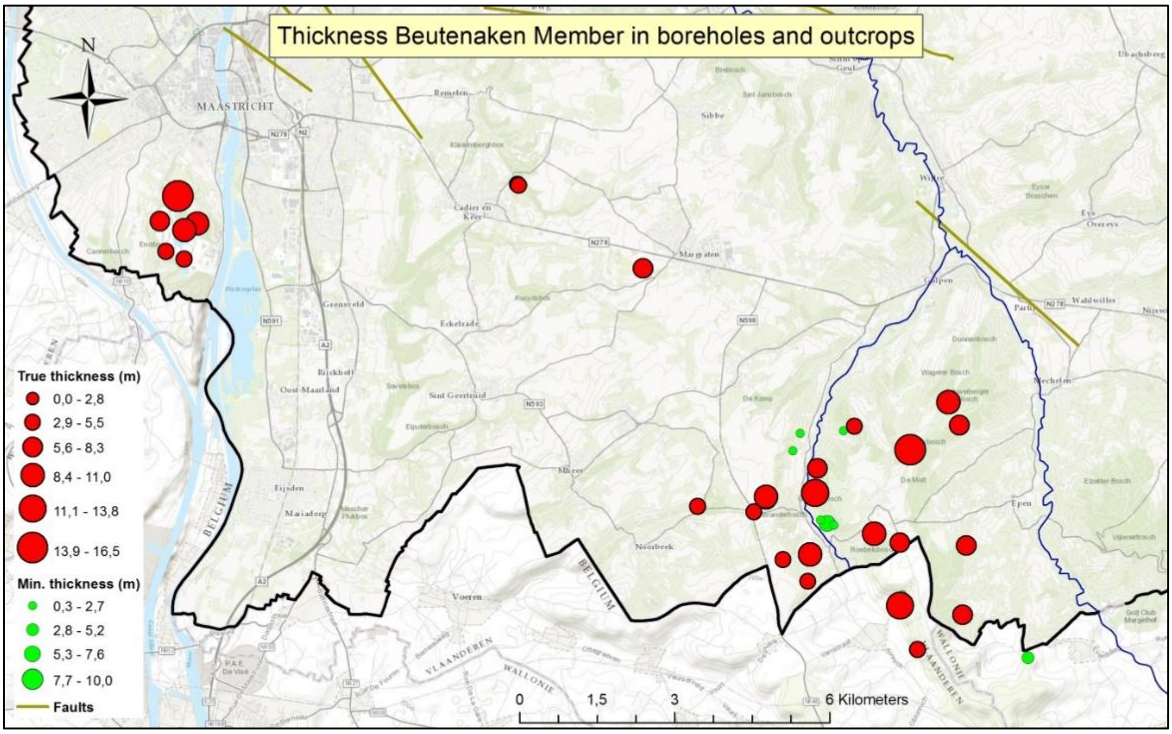

Fig. 4.5.2. Thickness of the Beutenaken Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Age
These sediments are placed in the Late Campanian based on belemnites (Van der Tuuk, 1980; Jagt et al., 1987; Keutgen et al., 1990).

Fossil Content
The sediments of the Beutenaken Member are characterized by specific benthic foraminifera, grouped as Zone B by Hofker (1966). This indicates that the sediments were formed in similar depositional environmental conditions. The profiles of Felder (2001) show relatively large numbers of pelecypoda and prismatic pelecypoda, as well as belemnites of which many are intact. No brachiopods were found in these sediments (Felder, 2001), while Felder et al. (2000) only found few remains.

Interpretation of depositional environment
The lack of brachiopods is an indication for relatively low-energy and fine-grained sedimentation (Felder, 2001). This is confirmed by the large abundance of prismatic pelecypoda, that require a soft substrate to burrow trough (mud or chalk ooze) (Felder, 2001). The amount of belemnites varies, however a relative high abundance of complete belemnite rostra was recognized (Felder, 2001). This indicates that the clasts have experienced no, or only short-distance transport. The strong increase of prismatic pelecypoda, relative to the underlying Zevenwegen Facies, indicates a relative deepening of the depositional environment. This is not in line with the relatively low $CaCO_3$ content which indicates a relative proximal depositional environment as the siliciclastic content is high. This however, can also be caused by increased erosion rates of the inverted Roer Valley Graben.

Although the distribution of the Beutenaken Member is limited, there is evidence that indicates that this member was present over a much broader area. Belemnites characteristic for this facies were recognized in burrows in the cemented top of the Zevenwegen Facies (Keutgen et al., 1990). This indicates a period of erosion after the deposition of the Beutenaken Facies unit. In most locations the Beutenaken Member has been eroded, south of Maastricht and around Slenaken the lower part of these sediments were preserved. This has been demonstrated by Felder (2001) who showed that in northwest Belgium, the Beutenaken Facies also comprises younger sediments as no erosion has occurred. Since this facies unit resembles a relatively deep depositional environment (offshore zone), deposition is more likely than non-deposition. This leads to the hypothesis of a severe erosional event after the deposition of the Beutenaken Member, presumably caused by a large marine regression.

4.6 Vijlen Facies

The Vijlen Member of Felder (1975) can be subdivided into two facies that gradually, both laterally and vertically, pass into each other.

Vijlen Facies A
A light-grey glauconitic chalk with small grey flint nodules. According to Bless et al. (1987) these are calcisiltites with a bioclast grain size of 30 to 45 µm. The flint content may reach 10% of the total volume in the upper part of this facies in the ENCI quarry (RGD, Rapport no. OP 5730). The glauconite concentration is highest at the base and gradually decreases upwards. In the southwest this facies contains little very fine-grained glauconite (Felder, 1983). The exact amount of glauconite is not known.

This facies is found dominantly in the western part of the study area. In the ENCI quarry it forms the upper part of the Vijlen Member, characterized by a relatively high carbonate content of 80 to 90%

(RGD, Rapport no. OP 5730). In Hallembaye sediments of this facies form the largest part of the Vijlen Member.

Vijlen Facies B
A yellowish to whitish glauconitic marl to marly chalk with quartz grains and locally coarse fossil-grit layers (Keutgen et al., 1990). The base is characterized by a high abundance of glauconite and fossils. It is especially enriched in belemnites in the so-called 'belemnites-cemeteries' (Fig. 4.6.1), which are local accumulations of belemnites alternated with hardgrounds (Felder, 1960). Next to glauconite and a high abundance of fossils, well- to poorly-rounded pebble layers and phosphorized fossil fragments are also present in this basal fossil-grit layer (Felder et al, 1994). It varies in thickness from tens of centimetres to several metres. The glauconite concentrations gradually decreases upwards but remains present throughout this facies.

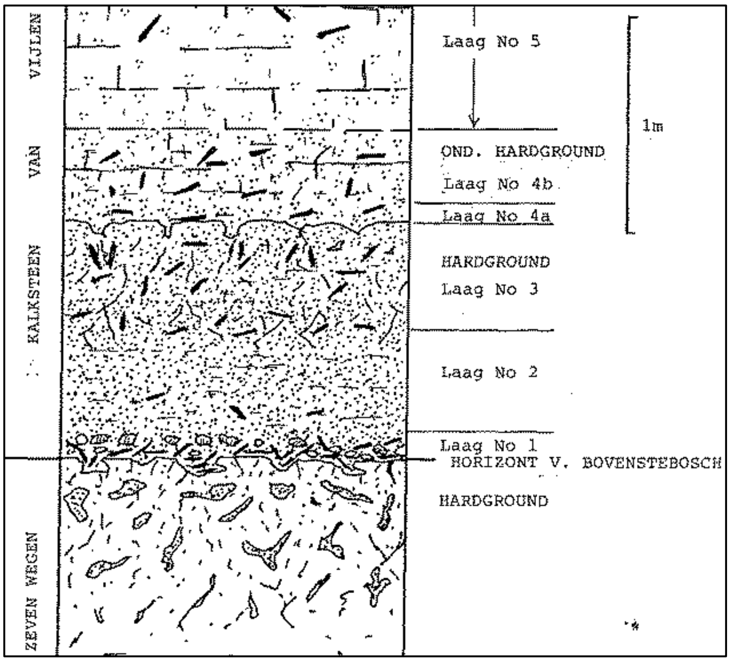

Fig. 4.6.1. The transition between the Zevenwegen Member and the Vijlen Member. The Belemnite cemetery' is well developed and multiple hardgrounds can be observed. Quarry 62D26, near Zevenwegen. (Sketch by Felder, W.M.)

This facies is found everywhere in the study area. In the west (ENCI quarry), the lower part of the Vijlen Member can be classified as the Vijlen Facies B. The $CaCO_3$ content ranges from 60 to 70%. Also around Hallembaye a marly glauconite-rich limestone is found at the base of the Vijlen member (Felder et al. 2001). The $CaCO_3$ content gradually increases from 75 to 90% (Felder, 1973). In the east, all sediments of the Vijlen Member can be classified as Vijlen Facies B. However, the top of these sediments consists of a limestone with marly intervals (Keutgen et al., 1990), which may be interpreted as a transition between both facies. The $CaCO_3$ content ranges from 50 to 80 % (Harten, 1972; Felder, 1973).

The distribution of the Vijlen Member is limited by the line Meerssen-Valkenburg-Eys-Bocholtz (Fig. 4.6.2). Towards the north the Vijlen Member thins out as is well visible in both profiles (Appendix). In two areas the Vijlen Member directly overlies Carboniferous rocks as is visualized in both profiles (see Appendix). It can also be observed in the thickness map (fig. 4.6.3) which shows larger thicknesses in these two areas. Around Eijsden this occurs in a zone of 2 km wide and 10 km long From Vaals to Nyswiller this occurs in a zone of 1 km wide and several kilometres long (see Appendix, profile 1). This zone is less clear in the thickness map (fig. 4.6.3) as incision of a creek (

Selzerbeek) has removed the upper part of the Vijlen Member. Striking is that the orientation of this creek is similar to the orientation of the previously described zone.

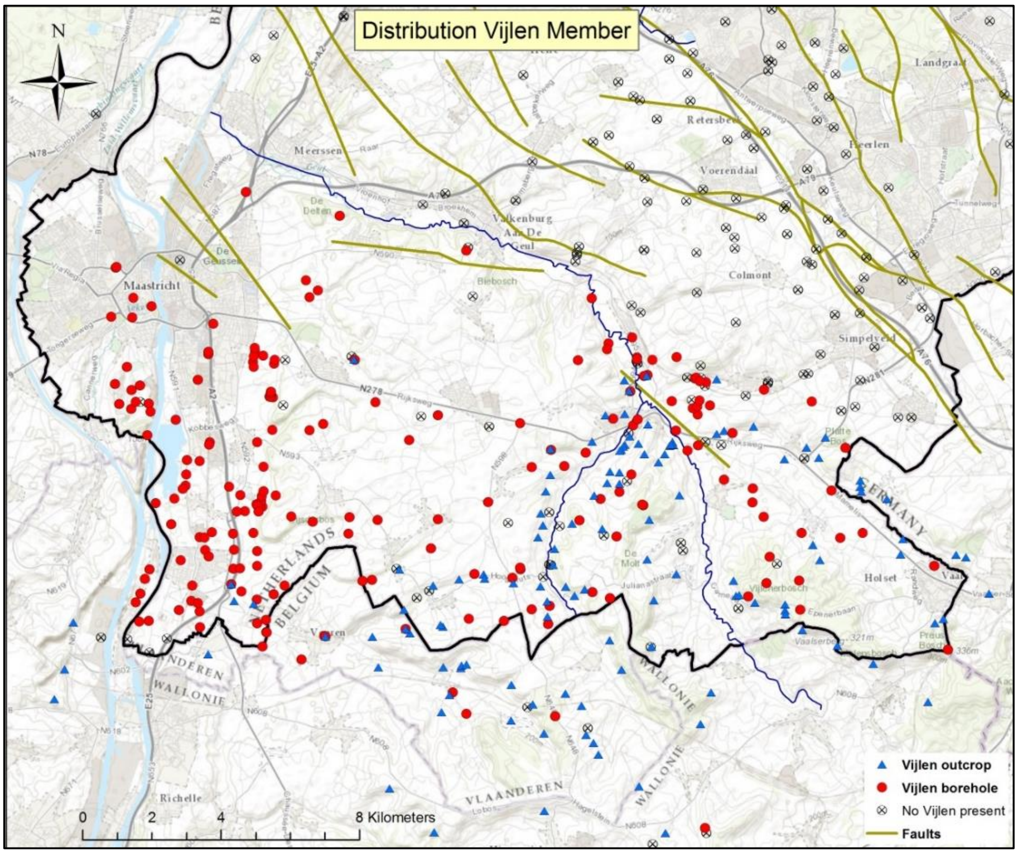

Fig. 4.6.2. Distribution of locations where the Vijlen Member is found.

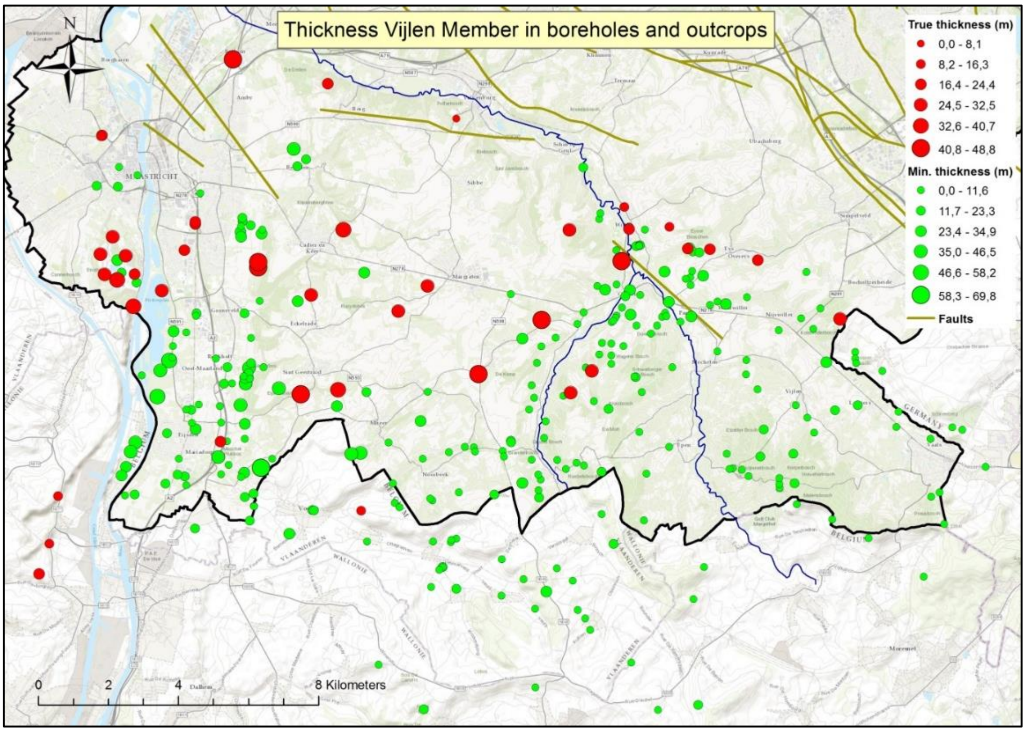

Fig. 4.6.3. Thickness of the Vijlen Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or

Pleistocene sediments).

Age
Keutgen and van der Tuuk (1990) placed the base of the Vijlen Member at an age of late Early Maastrichtian and its upper part in the earliest Late Maastrichtian, based on belemnites. This was confirmed by an ammonite study of Jagt and Kennedy (1994) and another belemnite study of Jagt et al. (1995).

Fossil content
The sediments of the Vijlen Member are characterized by specific benthic foraminifera, grouped as Zone C by Hofker (1966). This indicates that the sediments were formed during similar environmental conditions. Hofker (1966) did not recognize two facies in this unit, which may indicate that the depositional environment did not differentiate that much.

Felder et al. (1994) separated the Vijlen Member in seven intervals, named Vijlen 0 to 6, based on foraminifera abundances. Intervals 0 to 4 are only present in areas where the Vijlen Member is positioned on top of Carboniferous strata or where it reaches large thicknesses. Interval 5 and 6 are more common. According to Felder et al. (1994) the sediments in the southeast of the study area are characterized by major changes in the lateral and vertical composition of its bioclast and microfossil content. Molluscs and echinoderm (irregular) fragments are most dominant in these sediments (Felder, et al., 1994). The bioclast and microfossil content varies from south (outcrops Bovenste Bos , 62D-27 and Zeven Wegen, 62D-15) to north (outcrops Panhuis Vijlen, 62D-156; Jongensschool Vijlen, 62D-129; Mamelisserberg, 62D-130 and borehole B62D0168).

The base of this member (interval 0) contains a large abundance of belemnites (‘belemnite graveyard’) that can be found in the southeast of the study area. Towards the south belemnite clasts show an increased fragmentation. Higher in the sequence (intervals 1 to 6), intervals occur where mollusc bioclast assemblages are dominant in the north while echinoderm bioclast assemblages are dominant in the south. Starting from interval 2 the echinoderm bioclasts become gradually more dominant than the mollusc bioclasts. In the north of the study area alternating siliciclastic pebble horizons are only found together with dominantly mollusc bioclasts. These horizons thin out towards the south where echinoderm bioclasts are dominant.

Interpretation of depositional environment
The sediments of Vijlen Facies A contain a higher carbonate content than the sediments of Vijlen Facies B, which suggests that Facies Unit A was deposited in an environment with less siliciclastic supply. This can be interpreted as a shift from a relative proximal depositional environment to a relative distal depositional environment. The coarse fossil-grit layers at the base and throughout Facies Unit B indicate a relative shallow marine environment as well as high energy conditions.

This is confirmed by the presence of mollusc-dominated beds accompanied with pebble horizons, which are indicative for a shallow marine environment (Felder et al, 1994). Towards the south, the echinoderm-dominated sediments are more dominant, which is interpreted as an increased deepening of the depositional environment (Felder et al, 1994). This is confirmed by the presence of pebble horizons that thin towards the south, which suggests decreasing energy conditions as a result of deepening of the depositional environment. The dominance of irregular echinoderms suggests the presence of a soft substrate during deposition, which is another indication for a relatively deeper environment without wave influence (Brusca et al., 1990). This is also confirmed by Albers et al.

(1979) who found near-complete fish skeletons south of Vijlen. The diachronic northward withdrawal of belemnites goes parallel with a diachronic shift of the boundary between echinoderm-dominated and mollusc-dominated bioclasts towards the north. Felder et al. (1993) have interpreted this as a marine transgression towards the north, which is in the direction of the inverted Roer Valley Graben. This perfectly matches the facies classification which indicates a shift from a proximal environment to a more distal environment.

The facies model also indicates that lower energy conditions were reached in an earlier stage in the west (SW and NW) of the study area than in the east as shown by the dominance of the chalk facies. This can be interpreted as a relative deepening of the paleo-water depth towards the west. This could be explained by increased subsidence in the west, which is unlikely with the presence of the London-Brabant Massif. It is more likely that the eastern part of the study area experienced more uplift from the inverted Roer Valley Graben, and hence the transgression occurred with a slower pace resulting in more shallow marine environmental conditions than in the west.

Erosional features
Two zones were distinguished where a relative thick Vijlen Member directly overlies Carboniferous strata. This indicates erosion prior to deposition of the Vijlen Member. Felder et al. (1994) also recognized this as the sediment infill (interval 0 to 4) was limited to these two zones. The Aken, Vaals and lower part of the Gulpen Formation are eroded in two zones with a NW to SE orientation. After erosion, transgression resulted in renewed deposition and sediments classified as Vijlen Facies B were deposited. In the east of the study area fossil-grit layers are common, as well as accumulations of belemnites and siliciclastic grains. This can be interpreted as sediments from nearshore that are deposited by episodic flows towards the basin.

4.7 Lixhe Facies

The sediments of the Lixhe Member can be classified as a facies unit, called the Lixhe Facies. These are characterized by calcisiltites with a bioclast grainsize of 30 to 45 μm. Flint is present as flint nodule layers that increase in size upwards(Fig. 4.7.1).

The Lixhe Member can be found in most locations of the study area (Fig. 4.7.2). Towards the southeast it is absent. It may reach a thickness of up to 47.5 m (Fig. 4.7.3). On average, its thickness varies in between 20 to 30 m.

Fig. 4.7.2 Upper part of the Lixhe Member and the Lanaye Member. Photo at the Albert Canal during its widening and deepening in the 1980s. Also note that flint nodule layers are dipping towards the northwest (RGD).

The flint nodule layers in the west are laterally continuous and can be used for correlation. Towards the east planar layering disappears. The amount of flint reaches 15% of the total volume (RGD, Rapport no. OP5730-1). Measurements of $CaCO_3$ are only available for the west and southwest of the study area. Around Hallembaye (quarry Lixhe, 61H18) the lower part of this member contains 90 to 93% $CaCO_3$. Around Maastricht (boreholes ENCI quarry and boreholes 61F295 and 61F322) the $CaCO_3$ content varies between 80 and 90%. Towards the top, the $CaCO_3$ content gradually increases up to 96% (RGD, Rapport no. OP5730-1). The carbonate content varies from 80 to 90% in the east of the study area (Felder, 1973).

The base of this facies unit is formed in the east by a decimetre thick fossil-grit layer on top of a hardground. The fossil-grit layer contains a high concentration of glauconite, fossil fragments and siliciclastic grains ranging from very fine sand to gravel size (Horizon of Wahlwiller according to Felder, 1975). Towards the northeast (Wylre, Nyswiller, Eys), the grain size is further increasing and Carboniferous clasts and large quartz grains are incorporated in this layer. The Horizon of Wahlwiller is only found northeast of the line Wijlre – Margraten. Southwest of this line, this layer is not present nor is the hardground underneath it. The transition from the Vijlen Member to the Lixhe Member is formed by a flint-nodule layer in the southwest (Horizont of Lixhe according to Felder, 1975).

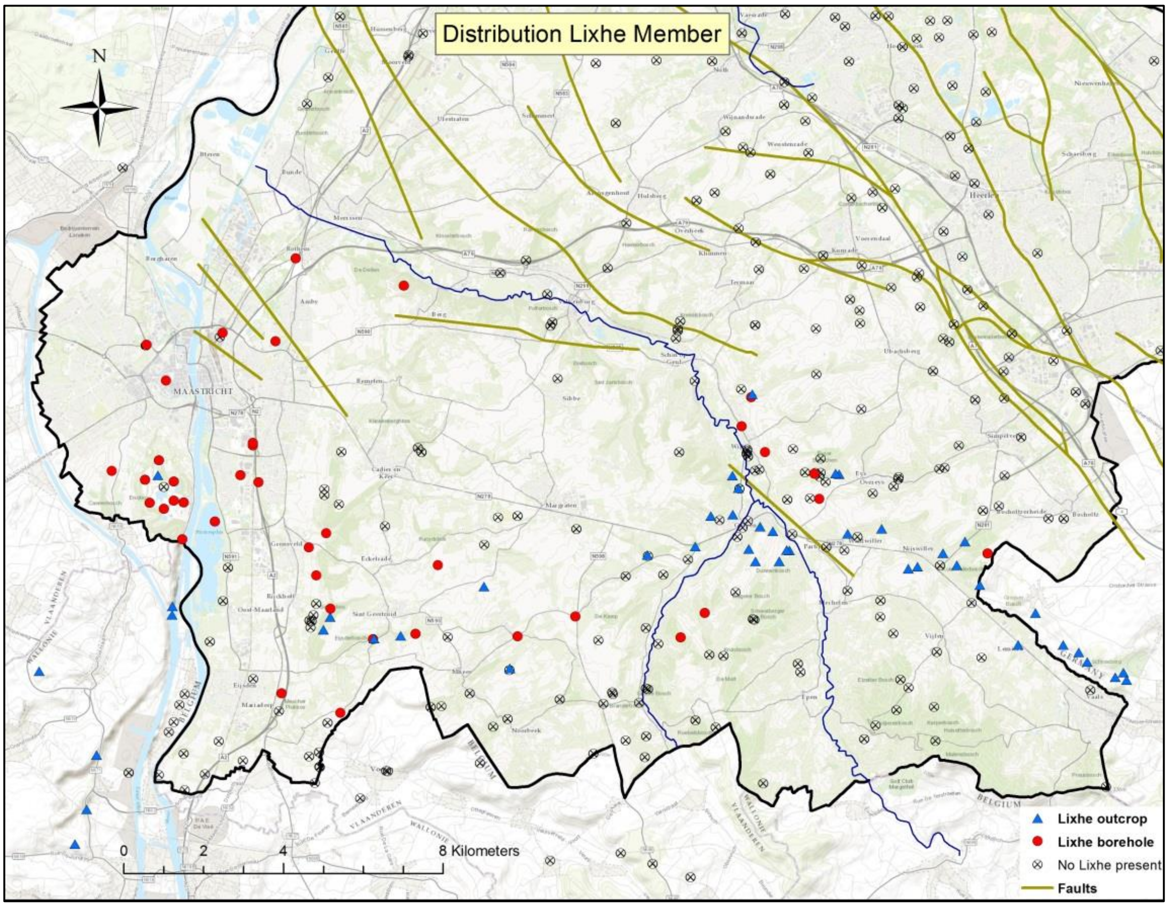

Fig. 4.7.2. Distribution of locations where the Lixhe Member is found.

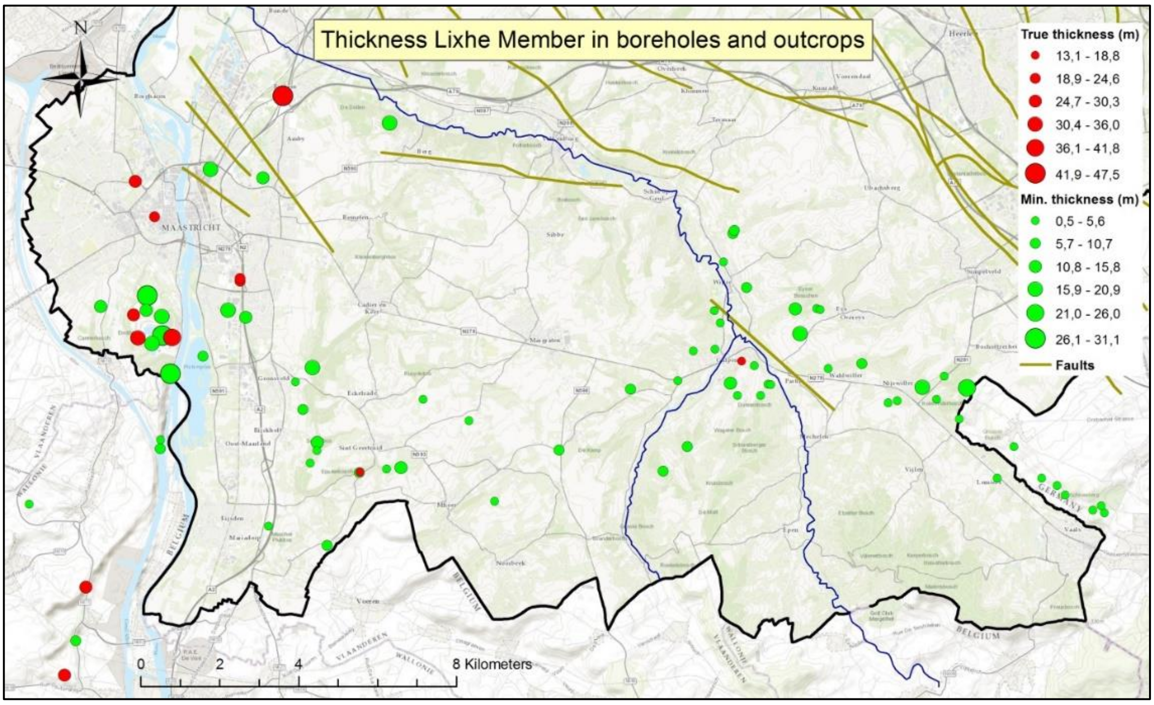

Fig. 4.7.3. Thickness of the Lixhe Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Age
Based on biostratigraphy, the sediments are the oldest strata in the Late Maastrichtian (Hardenbol et al., 1998).

Fossil content
The sediments of the Lixhe Member are characterized by specific benthic foraminifera, grouped as Zone D and E by Hofker (1966). This indicates that the sediments were formed during similar deposition environmental conditions.

According to Felder et al. (2000) the sediments of the Lixhe Member do not contain many macrofossils. The following observations were done in profiles of Felder (2001) that show a relative abundance of different fossil groups. Both profiles are from two quarries in Halembaye and Maastricht. The content of mesofossils shows for the first time a significant increase in numbers of crinoids which gradually increases towards the top. Also the number of prismatic pelecypoda slightly increases, while the number of pelecypoda is relatively low and decreases. The number of bryozoans slowly increases towards a maximum in the middle part of the sequence, followed by a gradual decrease. The relative abundance of brachiopoda is decreasing upwards in this facies unit. Small belemnoidae peaks are present in the lower part of the facies unit, but are significantly less abundant than in older facies.

Interpretation of depositional environment
Villain (1977) interpreted the sediments of the Lixhe Member to be deposited in a marine environment with oceanic influence, based on the presence of planktonic foraminifera and calcareous nannofossils. The water depth is assumed to have been between 80 and 150 m (Villain, 1977). The gradual increase in crinoids indicates a gradual deepening of the depositional environment as well as lower hydrodynamic energy conditions (Felder, 2001). A gradual deepening of the depositional environment can also be derived from the gradual increase in prismatic pelecypoda, which indicate the presence of chalk ooze during deposition (Felder, 2001)

The flint-nodule layers are formed by late diagenetic precipitation in zones of early diagenetic silica polymorphs (Zijlstra, 1995). The zones of early diagenetic precipitation occur around burrows and just below and parallel to the paleo sea floor (Bischoff and Sayles, 1972; Zijlstra, 1995). Lateral extensive flint nodule layers have formed as the late diagenetic precipitation of silica could continue as no new sediment was deposited (Zijlstra, 1995). The time between two flint nodule layers is 20 ky, in which Zijlstra (1995) recognized a precession related cause. The lateral continuity of the planar flint nodule layers indicate that the sediment surface was not reached by flows as it is not disturbed. This can be interpreted as a depositional environment below the storm wave base.

Towards the east of the study area the planar flint-nodule layers dissapear, and flint nodules are randomly distributed. This can be interpreted as a shallower depositional environment where successive storms reached the sediment surface regularly during a longer period, resulting in less pronounced planar flint layers. Also the fact that the basal fossil-grit layer, the Horizon van Wahlwiller (Felder, 1975) only developed in the east indicates relatively higher energy conditions than in the west. This suggests a relatively shallow-marine environment, above the storm wave base. The presence of Carboniferous organic clasts and coarse quartz grains, suggests that the inverted Roer Valley Graben was still a source of clastic sediments. The presence of a basal fossil-grit layer (Horizon van Wahlwiller) on top of a hardground may reflect a strong regression that caused erosion and a period of non-deposition resulting in the formation of a hardground. The regression is followed by a transgression which resulted in a basal fossil-grit layer as the paleo-water depth was presumably still less than the storm wave base. In the west of the study area, the paleo-water depth was already below the storm wave base during the regressional period, which explains the absence of fossil-grit layers or hardgrounds. The transition between the Vijlen and Lixhe Member is formed by a flint nodule layer (Horizon of Lixhe).

The aforementioned hypothesis is confirmed by the $CaCO_3$ content that slowly increases upwards, and westwards. This indicates a higher percentage of clastic grains in the east, suggesting a more proximal marine environment. The size of the flint nodules also increases upwards, which can be indicative of a deepening of the depositional environment since sedimentation rates decreased.

4.8 Lanaye Facies

The Lanaye Member is classified as a facies unit named the Lanaye Facies. The deposits are characterized by calcarenites with a bioclast grain size of 71 µm (Bless et al., 1987). The texture of the sediments of the Lanaye Member is described as packstone and grainstone (Bless et al., 1987). Zijlstra (1995) has described these sediments as a pure coccolithic wackestone to packstone with silt-sized bioclasts that show fining-upwards cycles that each show a flint nodule layer (Fig. 4.8.1). The grain-size cycles are bounded by wavy erosional surfaces. The largest part of a cycle is homogeneously bioturbated, but occasionally hummocky cross-lamination is preserved at the bottom (Zijlstra, 1995).

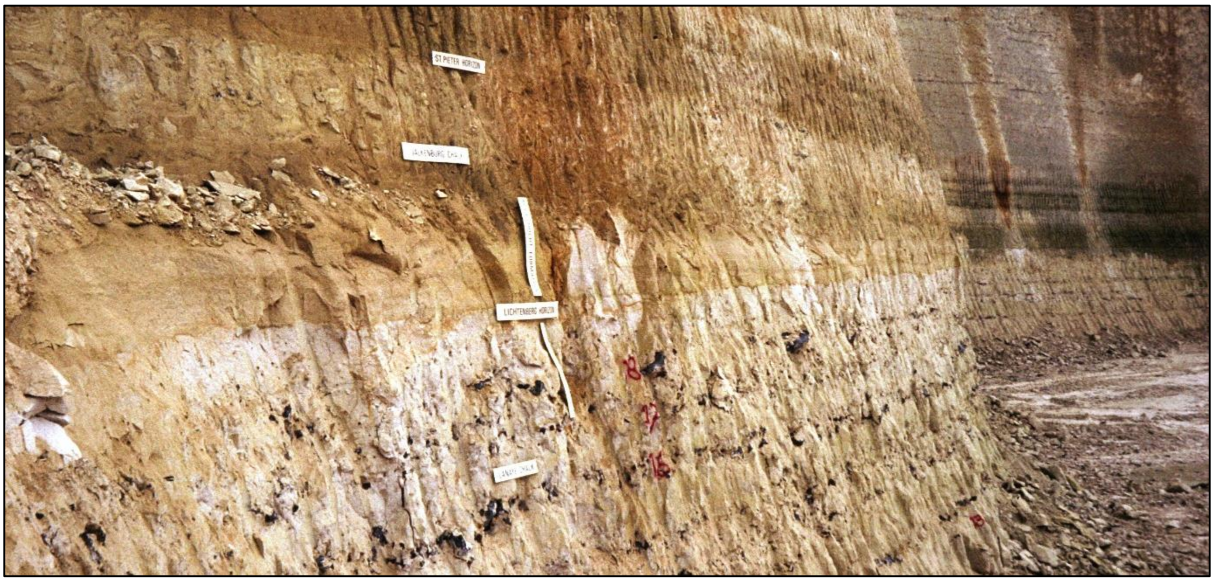

Fig. 4.8.1. The Lanaye Member with lateral flint nodule layers, it is covered by the Lichtenberg Horizon of the Maastricht Formation. Note the colour change from white to yellowish. Above the Horizon of Lichtenberg, the Valkenburg Member. ENCI quarry. (RGD)

The Lanaye Member is present everywhere in the study area, except in the southeast due to the uplift of the Ardennes (Fig. 4.8.2). The thickness varies from 3.5 to 21 m and increases towards the west. On average, the sediments reach a thickness of 14 to 20 m (Fig. 4.8.3).

The flint-nodule layers in the west of the study area are laterally continuous and can be used for correlation. Towards the east, planar layering disappears. The flint nodules reach 18 to 20% of the total volume (RGD, Rapport no. OP5730-1). According to Felder (1973) the carbonate content ranges from 80 to 90% in the southeast and 98% in the west of the study area. Around Aken the carbonate content is around 70% and around Valkenburg 90% (Felder et al., 2000). South of Maastricht (quarry

ENCI), the average $CaCO_3$ content of sediments is 97.4% (RGD, Rapport no. OP5730-1). The $CaCO_3$ content clearly increases upwards, from 94% at the base to 98% in the top.

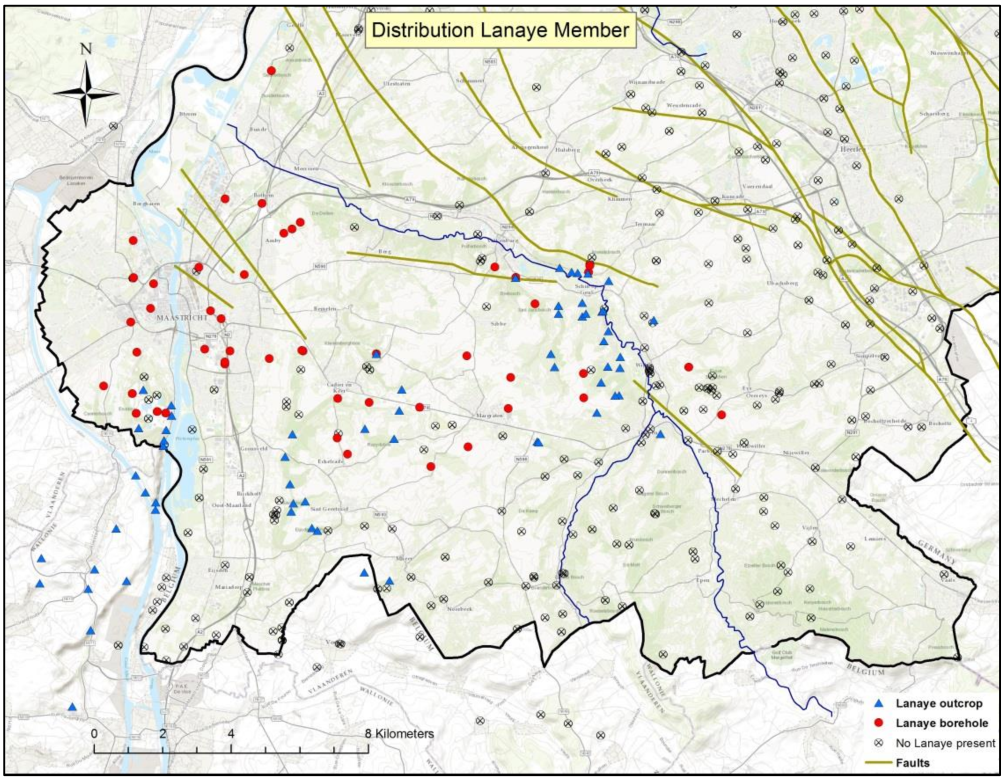

Fig. 4.8.2. Distribution of locations where the Lanaye Member is found

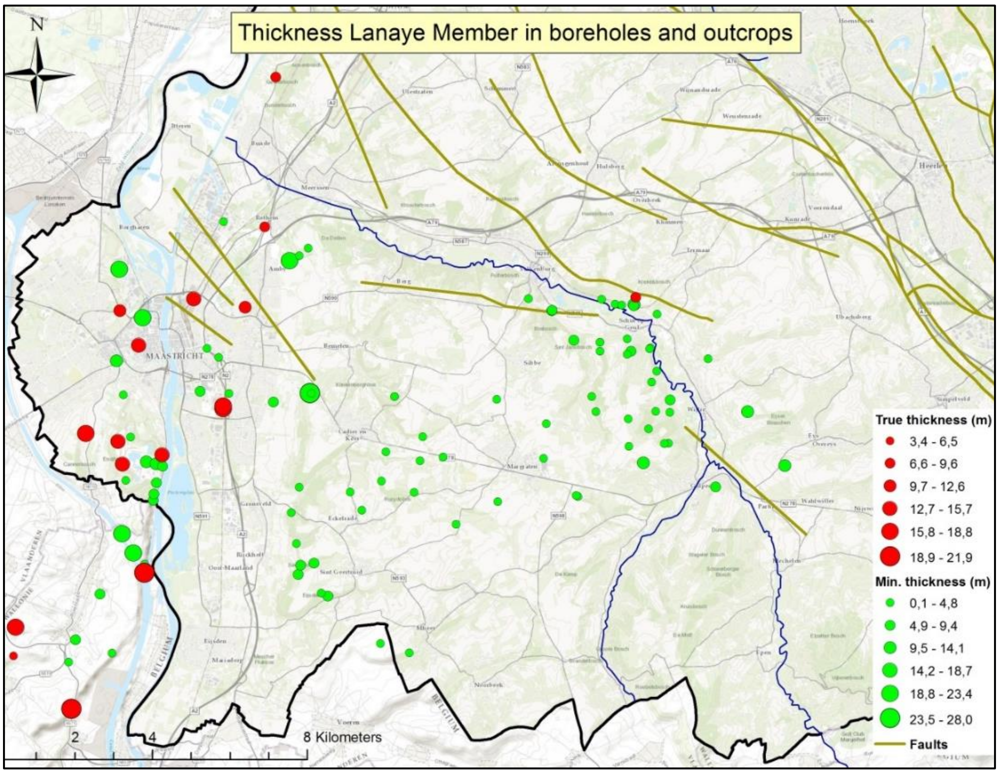

Fig. 4.8.3 Thickness of the Lanaye Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Age
Based on biostratigraphy, the sediments are of Late Maastrichtian age and cover the stratigraphically older Lixhe Member (Hardenbol et al., 1998).

Fossil content
The sediments of the Lanaye Member are characterized by specific benthic foraminifera, grouped as Zone F by Hofker (1966). This indicates that the sediments were formed during similar deposition environmental conditions.

The Lanaye Member contains many macrofossils. Typical for these sediments is the sea urchin *Echinocorys scutatus*, which increases in size with time (Felder et al., 2000). Thirteen different sea urchins have been described in the Lanaye member (Van der Ham et al., 1987). The following observations were done in profiles of Felder (2001) that show the relative abundance of different fossil groups in two quarries in Halembaye and Maastricht. The content of mesofossils shows for the first time a significant increase in numbers of crinoids and reaches a maximum just before the top of this member, followed by a steep drop. Also the number of prismatic pelecypoda slightly increases, while the number of pelecypoda is relatively low and decreases. The number of bryozoans slowly increases but remains relatively low. The relative abundance of brachiopoda is decreasing upwards in this facies unit. Belemnoidae disappear from the fossil content in this member (Felder, 2001), while Felder et al. (2000) states that rostra of belemnites are quite common.

Interpretation of depositional environment
The deposits of the Lanaye Member are interpreted to be deposited in a shallow marine environment with oceanic influence, and a water depth of 40 to 80 m (Villain, 1977). The presence of fining-upward cycles indicates a cyclicity in water depth or in hydrodynamic energy conditions. Zijlstra (1995) recognized a periodicity of 20 ky in the associated flint nodule layers. He also stated that this periodicity was caused by a cyclicity in storm intensity. During periods of maximum storm intensity the hydrodynamic energy conditions caused erosion and transport of sediments resulting in a period of non-deposition. When storm intensity was decreasing, deposition of sediment slowly continued again resulting in a lower concentration of early diagenetic silica and hence no preferential precipitation of late diagenetic flint occurred. The sediments on top of the erosional surface were deposited by storms and are relatively coarse-grained. As sedimentation continued, the hydrodynamic energy conditions during storms decreased, and so did the average grain size. After a storm intensity minimum, the hydrodynamic conditions increased again and the cycle repeated itself.

The difference with the underlying Lixhe Member is that the fining upward cycles are occasionally accompanied by hummocky cross-stratification. This may indicate two things: 1. The hummocky cross-stratification is destroyed by bioturbation in sediments of the Lixhe Member, which may indicate lower sedimentation rates. 2. Hummocky cross-stratification did not form in sediments of the Lixhe Member as they were deposited in a deeper depositional environment which was less influenced by storms.

The upward decrease of pelecypoda and crinoids indicates a relative shallowing of the depositional environment (Felder, 2001). This is confirmed by the strong increase in bioclast grain size which indicates deposition of biodetrital clasts during relatively higher energy conditions. The $CaCO_3$ content however, keeps increasing to a maximum of 98% in the west of the study area. This can be explained by a relatively large distance to the clastic source area. The gradual decrease in $CaCO_3$ content from 90% in Valkenburg to 70% in Aken indicates that the source area is located in the east as the clastic grain content increases. This was also recognized by Albers et al. (1979) and Felder et al. (2000) who indicated a facial difference between west and east.

Liebau (1978) interpreted these sediments to be deposited in the middle sublittoral zone and during sub-tropical temperatures. Influence of the Tethyan ocean to the south was also increasing, evidenced by the presence of equatorial ostracod species, mosasaurs and turtles (Bless, 1989; Jagt, 2010).

4.9 Maastricht Facies

The lower members of the Maastricht Formation (Valkenburg, Gronsveld, Schiepersberg and Emael Members) have a similar depositional origin and are classified as the Maastricht Facies. These sediments consists of calcarenites with flint nodules and fossil-grit layers of varying thicknesses. The mean grain size of bioclasts increases upwards from 90 to 120 µm (Bless et al., 1987). The texture of these sediments can be described as packstones and grainstones according to the Dunham classification. The geographical distribution is limited to the west and northwest of the study area, as Cenozoic uplift of the Ardennes eroded the strata to the southeast of the study area. Its total thickness ranges from 20 to 50 m.

In the west of the study area flint nodules are located in planar layers. South of Maastricht (ENCI quarry) the amount of flint ranges between 2 and 10%. The $CaCO_3$ content ranges from 92 to 96% for the Valkenburg, Gronsveld and Schiepersberg Members and from 96 to 98% for the Emael Member. Fossil-grit layers are present, but are barely visible and reach a thickness of only a few centimetres.

Towards the east and northeast the Maastricht Facies passes gradually, both laterally and vertically into another facies. The transition zone is called the Schaelsberg Facies by Felder (1975). In the Schaelsberg Facies flint nodule layers are less well developed and difficult to observe, instead flint nodules occur more randomly. Fossil-grit layers are more common, often several centimetres thick.

Valkenburg Member
The Valkenburg member in the west of the study area (Eben-Emael) consists of a 2 m thick calcarenite with two flint nodule layers. In the area of Maastricht (ENCI quarry) the member consists of a 3 m thick calcarenite with less organized planar flint nodules and a several centimetre thick basal fossil-grit layer with glauconite. It contains a $CaCO_3$ content of 94% at the base and 96% at the top (Felder, 1975). The flint nodules make up 5 to 8% of the total volume (Felder et al., 2000). The basal fossil-grit layer is a coarse-grained phosphatic-glauconitic-pyritic bioclastic sand on top of an undulating hardground in the top of the Gulpen Formation (Zijlstra, 1995). The transition between the undulating top surface of the hardground and the base of the fossil-grit layer, is named the Lichtenberg Horizon (Felder, 1975). The basal fossil-grit layer varies in thickness, from several centimetres to 20 cm. It is known for the many broken fossil clasts, characterized by shark teeth, coprolithes, belemnites and oysters (Felder, 1975). Also pebbles of intraformational limestone can be observed. Van Harten (1972) showed that the base is characterized by an increased concentration of coarse-grained quartz. This basal fossil-grit layer was only exposed south of Maastricht (quarry ENCI), it is unknown how it is developed further eastward since it is only known from outcrops. Zijlstra (1995) recognized three fining upwards cycles with the highest glauconite concentration at the base of each cycle.

The thickness of the Valkenburg Member increases to 25 m in Bemelen and Cadier en Keer where it contains a $CaCO_3$ content of 86% in the lower part and reaches up to 92% in the upper part (Felder et al., 2000). In Valkenburg this member reaches a thickness of 35 m and is present as a glauconitic calcarenite with irregular layers and lenses of cemented beds (Fig. 4.9.2). The glauconite concentration decreases from a glauconite-rich layer of 8 m thick at the base, to no glauconite in the top (Felder et al., 2000). Only in the top small flint nodules can be observed. The $CaCO_3$ content is on average 75% in the lower part and up to 90% in the upper part (Felder et al., 2000).

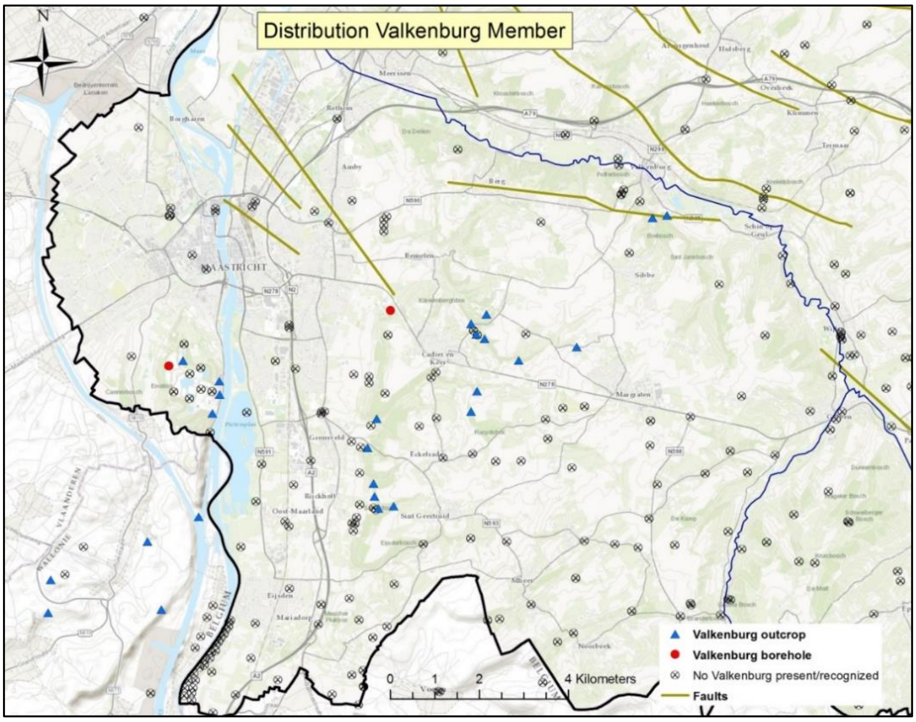

Fig. 4.9.1. Distribution of locations where the Valkenburg Member is found.

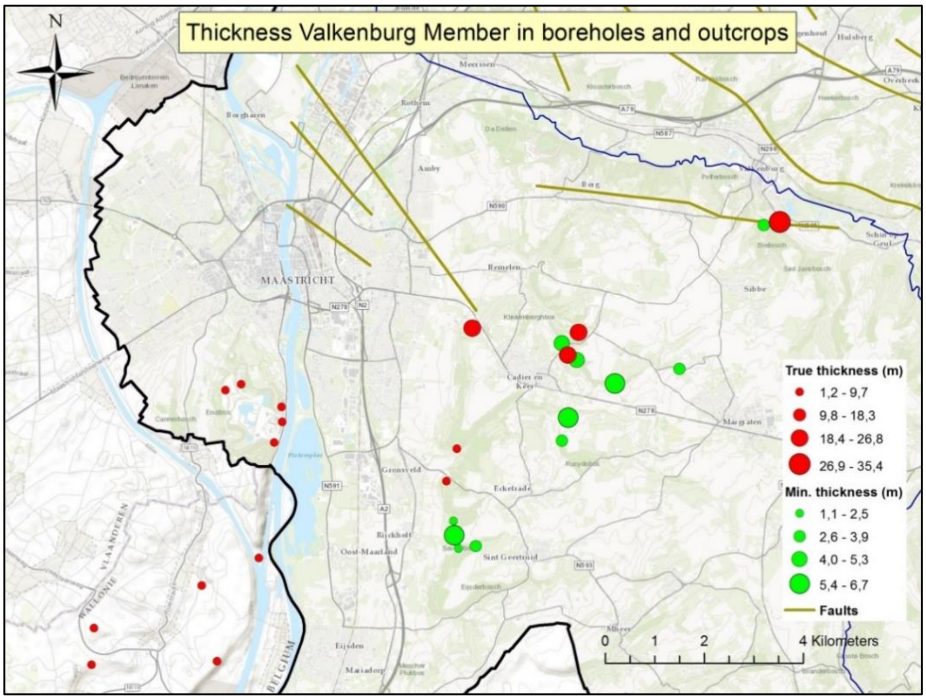

Fig. 4.9.2. Thickness of the Valkenburg Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Gronsveld Member

The Gronsveld Member consists of a calcarenite with flint nodules of varying sizes and small fossil-grit lenses. Its distribution is limited to the northwest of the study area (Fig. 4.9.3), as is the case of all members of the Lower Maastricht Formation. South of Maastricht (ENCI quarry) this member was well exposed and most research was done at this location. The base of the member is made up of a calcarenite alternated with thin fossil-grit lenses on top of an underlying undulating hardground, this transition is called the Horizon of St. Pieter (Felder, 1975). Above, Zijlstra (1995) recognized fining-upward cycles coarse-grained phosphatic-glauconitic-pyritic bioclastic sand at the base. The base of each cycle is characterized by wavy lamination that gradually disappears towards the top of each cycle where a homogeneously bioturbated calcarenite is found. This cyclicity is only found in the lower part of the Gronsveld Member, the upper part consists of a well sorted bioclastic sand with large-scale hummocky cross stratification (Zijlstra, 1995). Its thickness ranges from 4 to 10 m (Fig. 4.9.4). The $CaCO_3$ content varies between 92 to 97% and shows a strong decrease at the base of every fining-upwards cycle (Felder, 1975). The flint content ranges from 5 to 10% of the total volume in the west of the study area (Felder et al., 2000).

Further south (Eben-Emael), this member developed as a calcarenite with two distinct flint nodule layers. According to Felder (1975) the top of this member is represented by the top of the upper flint nodule layer. At Gronsveld (outcrop 62A-159), southeast of Maastricht, the Gronsveld Member reaches a thickness of 14 m. It consists of a calcarenite with planar flint-nodule layers that are increasing in size upwards. Above the Horizon of St. Pieter, laminated fossil-grit layers formed. In Valkenburg, further towards the northeast, outcrop 62A-294 shows a soft calcarenite with randomly-spread small flint nodules on top of the laminated basal fossil-grit layers.

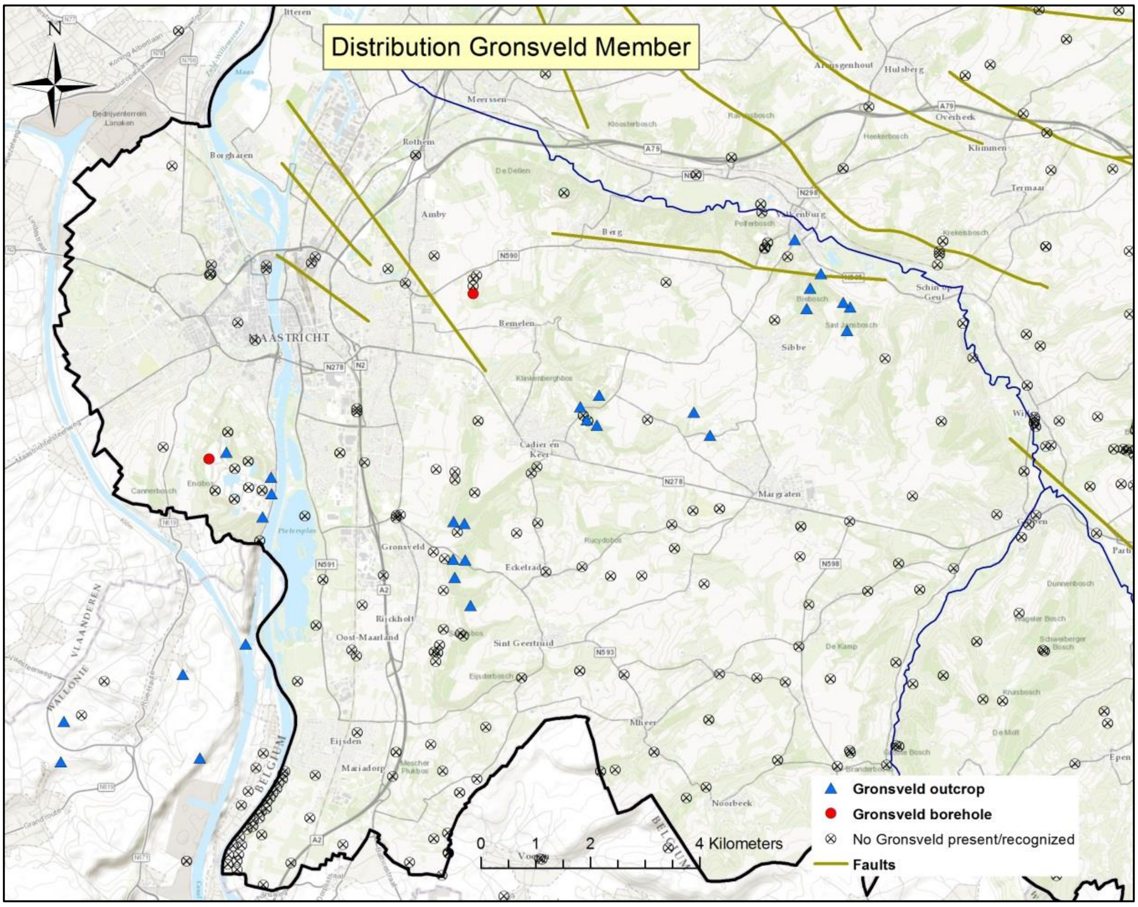

Fig. 4.9.3. Distribution of locations where the Gronsveld Member is found.

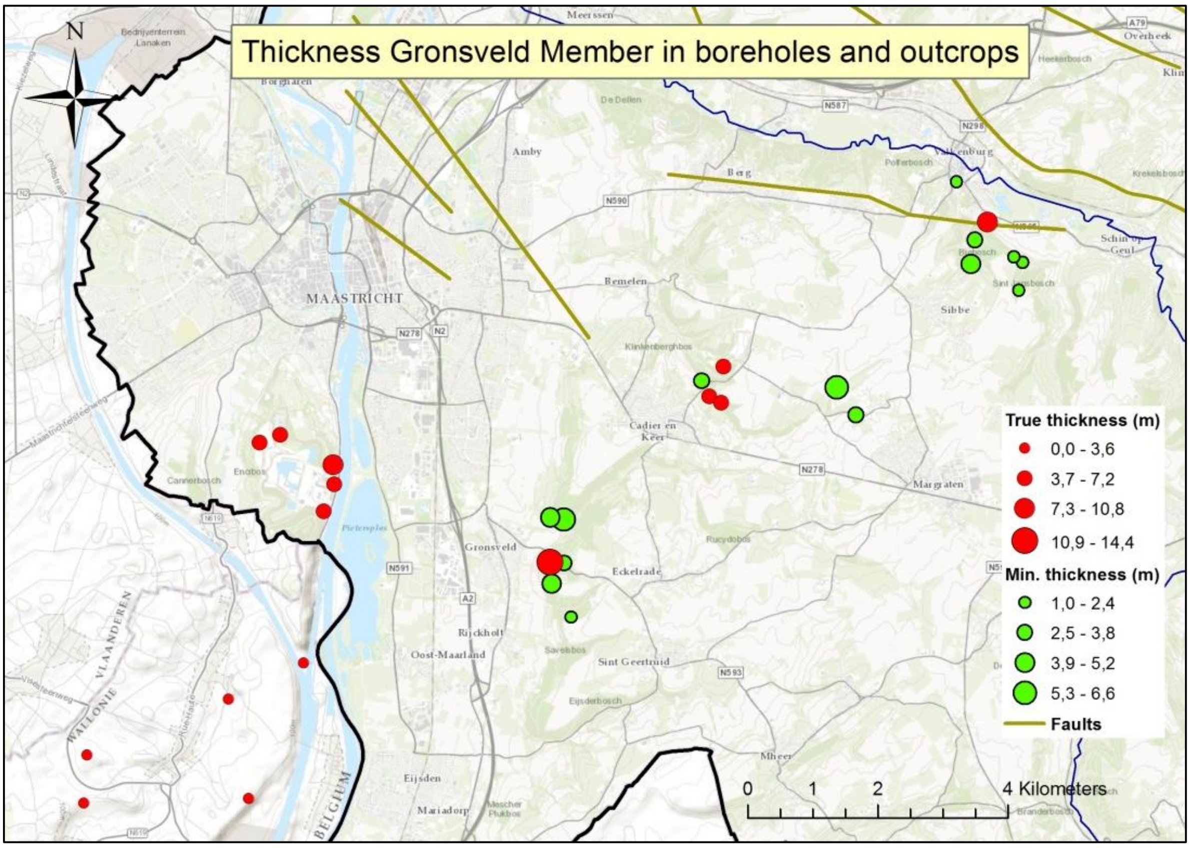

Fig. 4.9.4. Thickness of the Gronsveld Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Schiepersberg Member

The Schiepersberg Member in the Maastricht and Cadier en Keer area, consists of a calcarenite with relatively large flint nodules distributed in planar layers (Fig. 4.9.5 and Fig. 4.9.6). Towards the east the flint nodule layers disappear and the flint nodules are more randomly distributed (quarry Blom, 62A19, southeast of Curfs quarry). The distribution of the member is limited to the northwest of the study area as older strata are exposed due to the Cenozoic uplift of the Ardennes (Fig. 4.9.7).The lower part consists of laminated fossil-grit layers on top of an irregular cemented layer below, the interval in between is called the Horizon of Schiepersberg. The thickness of this member ranges from 3 to 6 m (Fig. 4.9.8). The average $CaCO_3$ content is 96% in the ENCI quarry, south of Maastricht. Further south (Eben-Emael) this member is absent.

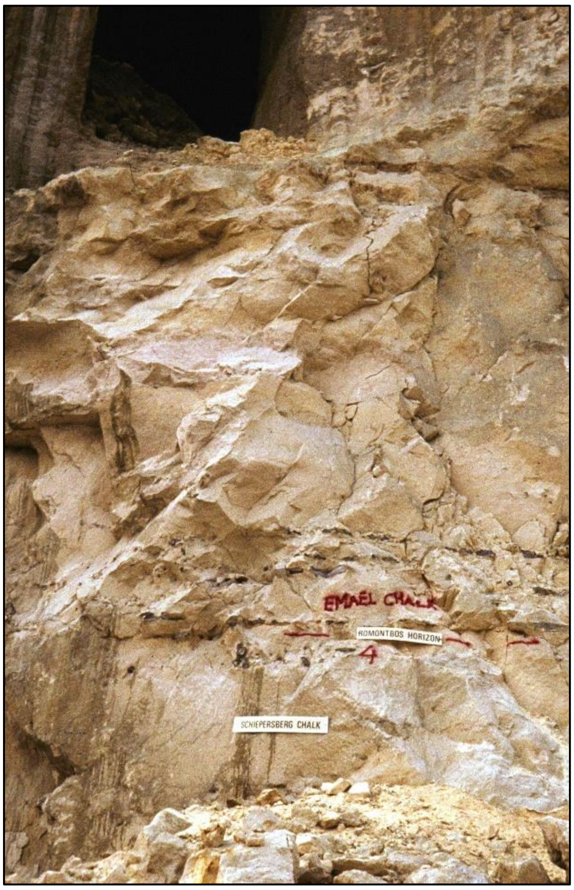

Fig. 4.9.5, left photo: The upper part of the Gronsveld Member, the Schiepersberg Member and the lower part of the Emael Member with the Horizon of Romontbos at its base. Quarry 't Rooth. Fig. 4.9.6, right photo: The Emael Member with the Horizon of Romontbos at its base and a cemented layer in the top (Horizon of Laumont). The indurated layer below the mine tunnel is the top of the Horizon of Laumont. Quarry ENCI. (RGD)

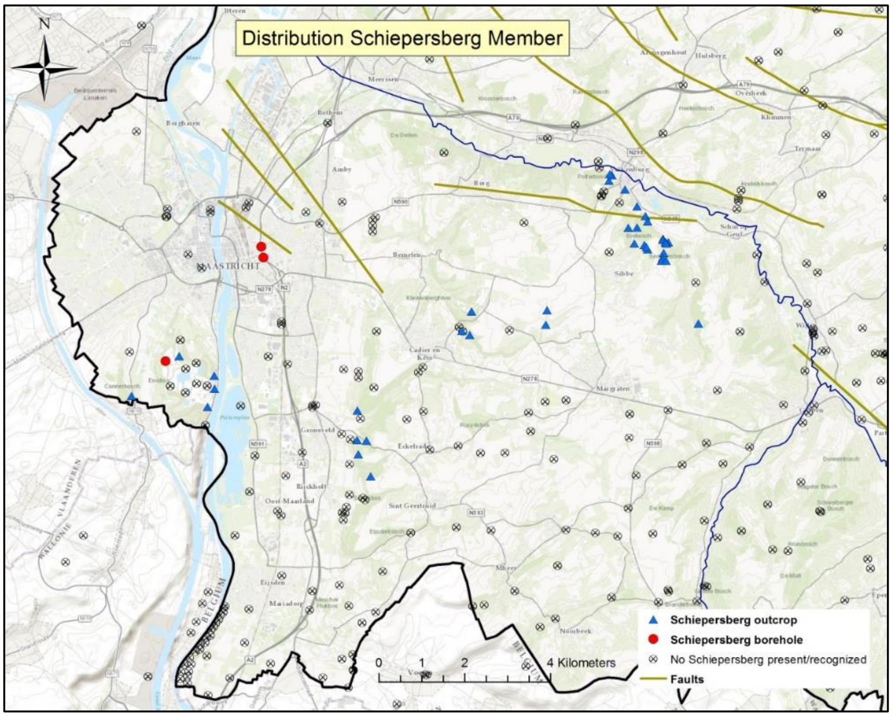

Fig. 4.9.7 Distribution of locations where the Schiepersberg Member is found.

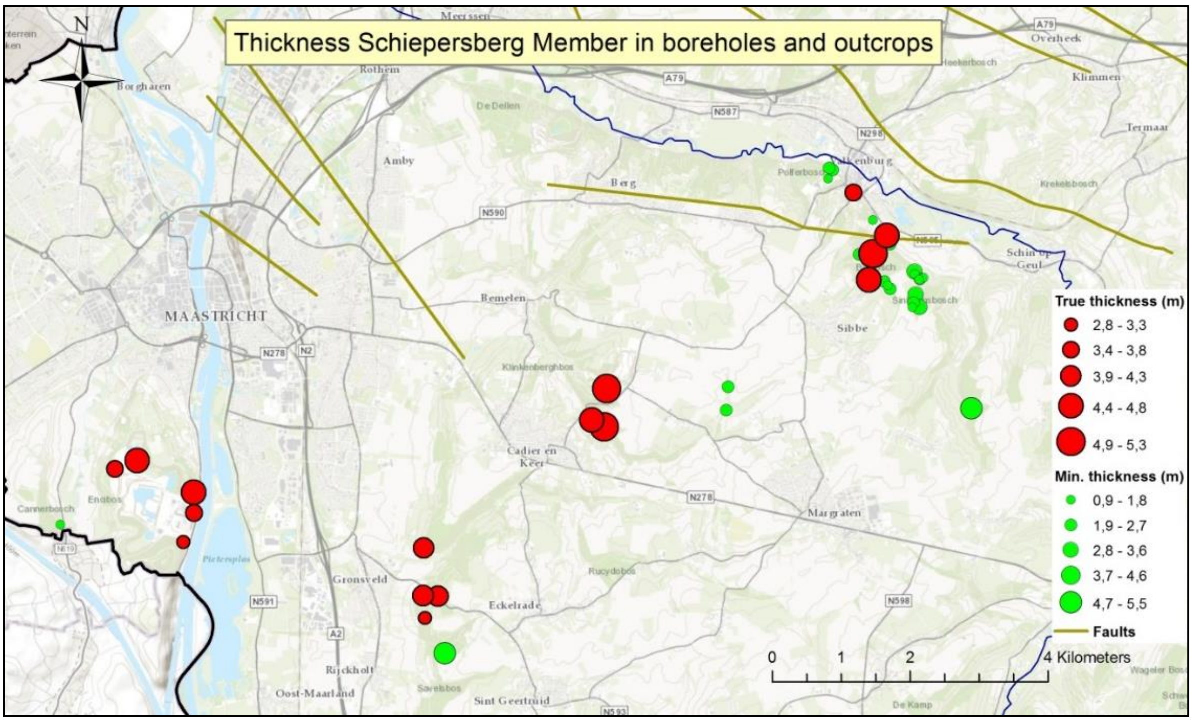

Fig. 4.9.8. Thickness of the Schiepersberg Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Emael Member

This member consists of a calcarenite that can be separated in two parts. The lower part consists of a basal 2 m thick layer with characteristic flint nodules that are pipe-shaped and orientated horizontally (Photo 4.9.9). Above, homogeneously bioturbated calcarenite with a varying thickness is found with in the top a hardground called the Horizon of Lava by Felder (1975). The lower part varies in thickness, from 3 to 4 m west of the river Maas and 5 to 7 m around Cadier en Keer, to 2.5 m around Valkenburg. The upper part, with a thickness of 3 to 4 m, of the Emael Member lacks flint nodules and contains large fossil-grit lenses and nodular cemented beds that are discontinuous.

The distribution of the Emael Member is limited to the northwest of the study area (Fig. 4.9.10). The upper part of the Emael Member is only found west of the river Maas (Eben-Emael and ENCI quarry). East of the river Maas, the upper part is absent as well as the Horizon of Lava.

Photo. 4.9.9. Lower part of the Emael Member with characteristic thick pipe-shaped flint nodule. On top lamination is visible. Quarry 'T Rooth (RGD)

The base of this member contains a coarse-grained fossil-grit layer formed on top of a cemented layer, the transition is called the Horizon of Romontbos (Felder, 1975). This fossil-grit layer is found everywhere in the study area except around Eben-Emael. Around Eben-Emael, a flint nodule layer below the Horizon of Romontbos is far better developed than in any other location (Felder, 1975). The top of the Emael Member is formed by the Horizon of Laumont (Felder, 1975). This is the top of a well-cemented hardground on the east side of the river Maas. On the west side, this hardground is absent or not well developed. In the ENCI quarry, a cemented layer with some burrows can be observed. While around Eben-Emael, no cementation at all occurred and the transition to the overlying sediments is formed by a thick flint nodule layer. The total thickness of this member varies from 2 to 9 m, the largest thicknesses are found on the west side of the river Maas (Fig. 4.9.11). The $CaCO_3$ content is 97% in the area south of Maastricht (ENCI quarry).

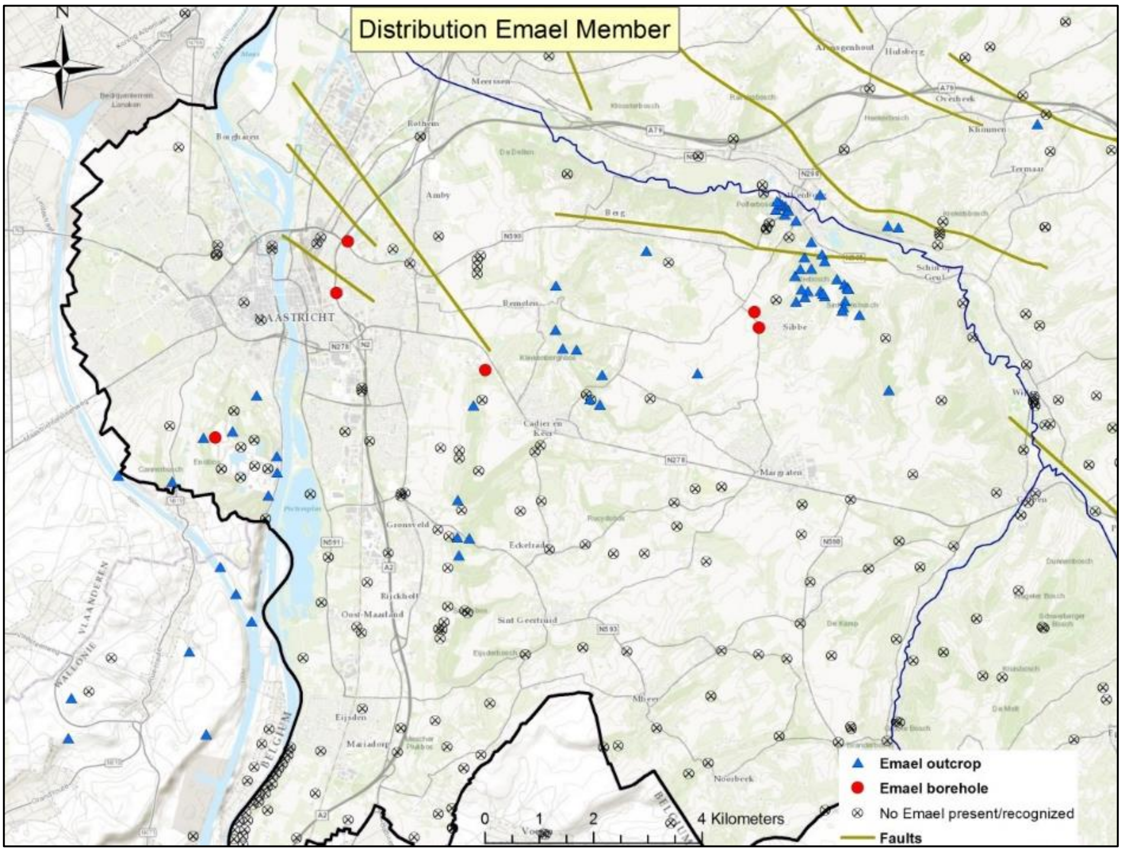

Fig. 4.9.10. Distribution of locations where the Emael Member is found.

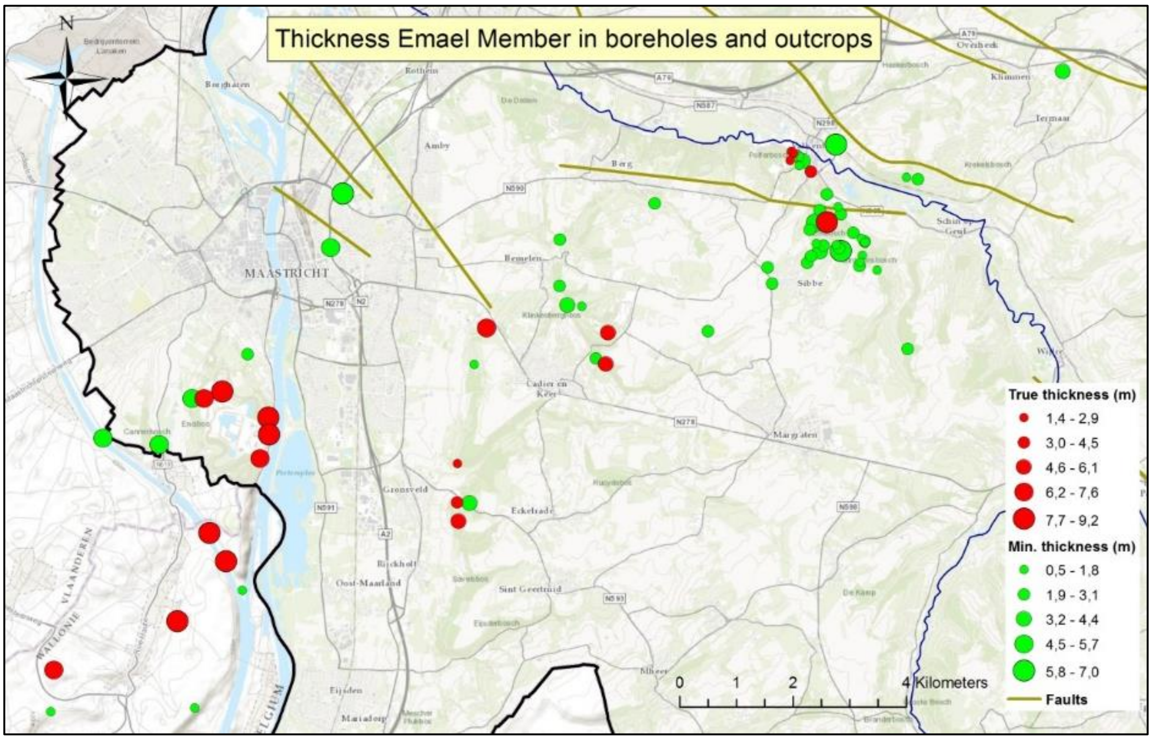

Fig. 4.9.11. Thickness of the Emael Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Fossil content

The Valkenburg, Gronsveld, Schiepersberg and Emael Members correspond with several foraminiferal zones, distinguished by Hofker (1966). In the west of the study area, these are Zones G, H and I (Fig. 4.9.12). It appeared impossible to correlate the base or top of different members with specific foraminiferal zones. This suggests that lithology and benthic foraminera cannot be linked in great detail. Hofker also recognized a distinct foraminiferal assemblage in the area of Valkenburg, called the Schaelsberg Chalk (Hofker, 1966). This facies shows some resemblance with zone H but contains several other species that are far more abundant than in other zones (Hofker, 1966). This zone was also recognized in the area of Maastricht, where it forms a facies in between sediments belonging to zone H (Fig. 4.9.12). In Valkenburg the Schaelsberg Facies forms the base of the Maastricht Formation and is later replaced by Zone H. Towards the northeast the Maastricht facies gradually passes into the Kunrader Facies where the benthic foraminifera assemblages are different.

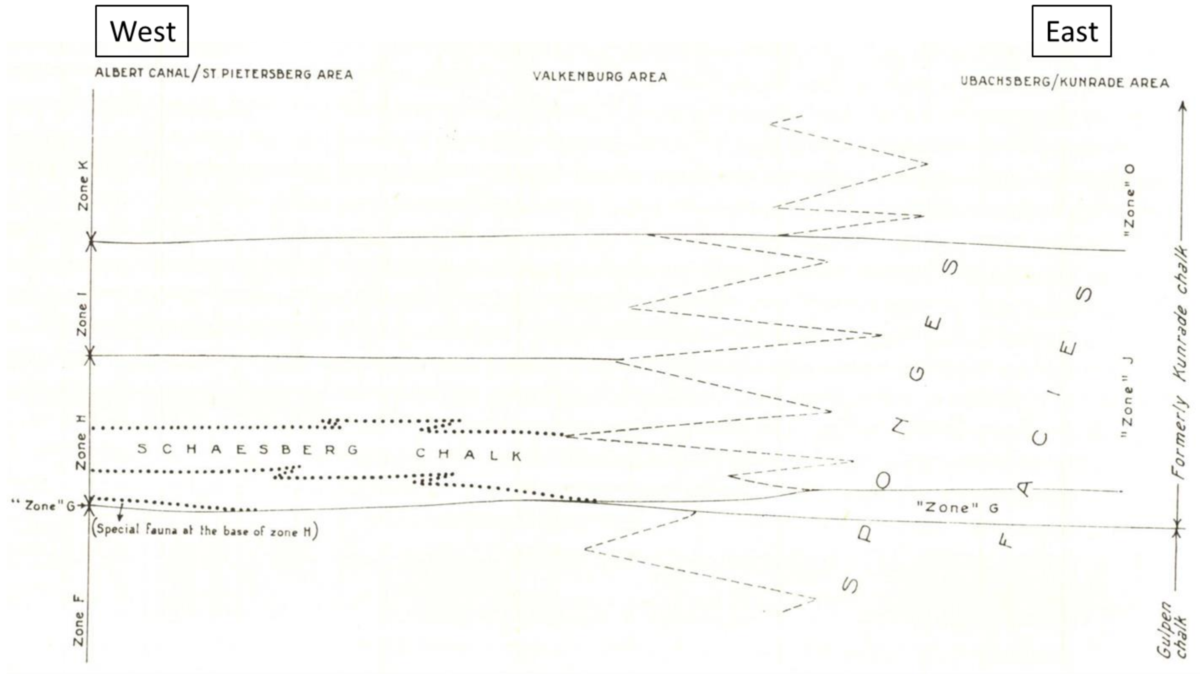

Fig. 4.9.12. Distribution of the different zones of Hofker (1966) in the study area and their interrelation (Modified after Hofker, 1966).

Observations by Felder (2001) show relative abundances of different fossil groups. The observations come from four different locations: outcrop 62A-163 in Gronsveld, outcrop 61H-36, quarry ENCI 61F-19 south of Maastricht, borehole 62F-296 in Maastricht. The number of crinoids is far less than in the Gulpen Formation and gradually decreases upwards in the Maastricht Facies. The number of serpulids is increasing in the sequence, it becomes the most abundant fossil group in the sediments with *Sclerostyla regia* as characteristic species. Numbers of prismatic pelecypoda and pelecypoda are relatively low, but are still found. Towards the top of the Emael Member, prismatic pelecypoda disappear and pelecypoda numbers are only few percent of the total. The number of bryozoa is also rapidly increasing towards a maximum at the base of the Emael Member, followed by a decrease.

Interpretation of depositional environment

One of the most important observations is the distinct presence of a flint-nodule layers that gradually disappears towards the east. This is accompanied by a gradual increase in cementation, as well as an increase in thickness of fossil-grit layers. At the ENCI quarry it forms a clear cycle as described in the previous chapter (Common characteristics). However, to the south (Hallembaye), it appears that no cementation occurred in the top layer neither did a fossil-grit layer form at the base. The flint nodules are larger in size and more pronounced as a layer. The flint-nodule layers are formed by late diagenetic precipitation in zones of early diagenetic silica polymorphs (Zijlstra, 1995). The zones of early diagenetic precipitation occur around burrows and just below and parallel to the paleo sea floor (Bischoff and Sayles, 1972; Zijlstra, 1995). Lateral extensive flint nodule layers have formed as the late diagenetic precipitation of silica could continue as no new sediment was deposited (Zijlstra, 1995). The lateral continuity of the planar flint nodule layers indicate that the sediment surface was not reached by flows as it is not disturbed. This can be interpreted as a depositional environment below the storm wave base, and a relative deepening towards the south and west. This is in agreement with the disappearance of fossil-grit layers towards the south and west.

The presence of lamination or cross bedding in fossil-grit layers at the base of each cycle is related to a cyclicity in storm intensity (Zijlstra, 1995). This 20 ky cycle resulted in a period of non-deposition due to erosion by storm waves at an intensity maximum. After an intensity maximum, the storm wave base gradually shallowed resulting in a decrease in energy conditions during storms. Previously deposited storm sediments, fossil-grit layers with occasionally lamination or cross-bedding, were not eroded anymore and deposition rates increased again.

The Emael Member can be separated in a lower and an upper part, separated by a hardground (Horizon of Lava). East of the river Maas, the upper part may be: 1. Not recognized since it is present as another facies; or 2. Eroded. The lower part is indicative for a relatively deep shallow-marine environment due to the presence of flint, the lack of fossil-grit layers (except the basal one) and the homogeneous soft appearance. The upper part, with thick fossil-grit layers and nodular cemented discontinuous layers is indicative for a relatively shallow depositional environment where the sediments were (periodically) reworked by waves (Inden et al., 1983). Hypothesis one suggests that the upper part is present in the northeast but in a different facies, this implies that the depositional environment deepened. This is contradictory with the fact that clear flint-nodule layers are disappearing and fossil-grit layers are thickening towards the northeast, which indicates that the depositional environment is shallowing towards the northeast. Hypothesis two is that the depositional environment in the northeast was relatively shallower as well as subject to higher hydrodynamic energy conditions. This probably led to erosion and periods of non-deposition. While in the west the depositional environment was deeper and the sediments were partly cemented and alternated with fossil-grit layers, but not eroded. This is in agreement with the observation that the top of the Emael Member only shows a hardground east of the river Maas, which indicates a period of non-deposition (Zijlstra, 1995).

Villain (1977) interpreted the depositional environment as a shallow-marine environment with a water depth of 20 to 40 m, without oceanic influence. Regional currents transported bioclasts over large distances. Liebau (1978) interpreted these sediments to be deposited in the middle sublittoral zone with sub-tropical temperatures, characterized by the occurrence of sea grass communities. The presence of cross bedding, lamination and fossil-grit layers shows that the paleo-sea bed was reworked periodically by waves. This can be interpreted as a depositional environment in between the fair weather wave base and the storm wave base.

4.10 Kunrade Facies

The Kunrade Facies is the lower part of the Maastricht Formation in the east of the study area. In other words: it is the coeval equivalent of the Maastricht Facies in the west of the study area . The sediments of this facies consists of a cyclic alternation of light grey hard and soft, calcarenites with a bioclast grain size of 105 µm (Bless et al., 1987). The hard beds are cemented by micrite. The alternate hard and soft beds are not continuous, they extend often only for several tens of metres (Fig. 4.10.1). Siliciclastic grains are relatively abundant: pebbles of shale and coal, quartz grains, glauconite and quartzite. The upper part of this unit may contain some small flint nodules.

Fig. 4.10.1. Sequence of sediments classified as Kunrade Facies at Kunderberg. Vertical scale 6 to 10 m. (From: RGD)

Hard beds have a $CaCO_3$ content of 92 to 98%, measured in ten locations throughout the area where sediments belonging to the Kunrader Facies were found (Kuyl, 1980). Relative soft beds have a $CaCO_3$ content of 69 to 89% measured in five locations. In Bocholtz, at the border with Germany, sediments are characterized by an average of 80% of hard beds and only 20% of soft beds. Towards the west, the percentage of hard beds decreases and soft beds become gradually more dominant.

The Kunrader Facies is only found in the northeast of the study area, south of the Benzenrade Fault (Fig. 4.10.2). The distribution map and the cross-sections (Appendix. Profile 1 and 2) show a strongly decreased thickness for these sediments north of the Benzenrade Fault. The total thickness of the Kunrade Facies is 50 to 60 m. However, it is often difficult to pinpoint the exact transition from the Zandig Krijt van Benzenrade Facies to the Kunrader Facies.

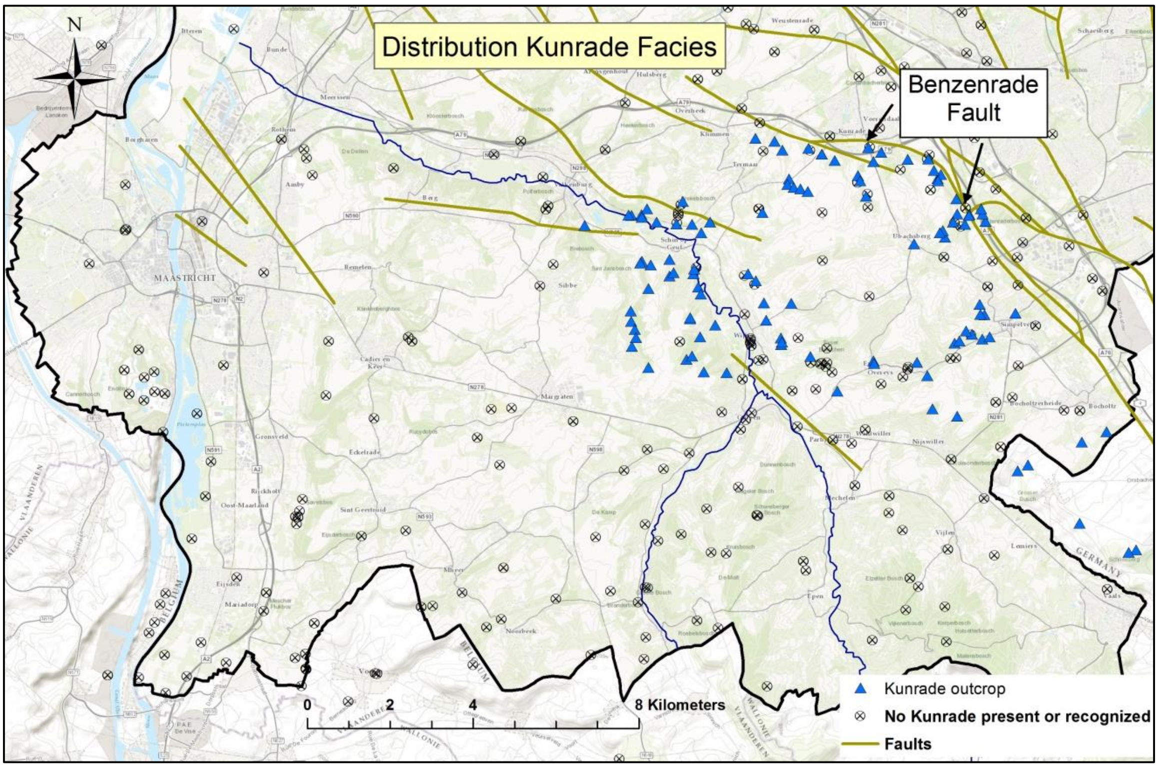

Fig. 4.10.2. Distribution of locations where the Kunrade Facies is found.

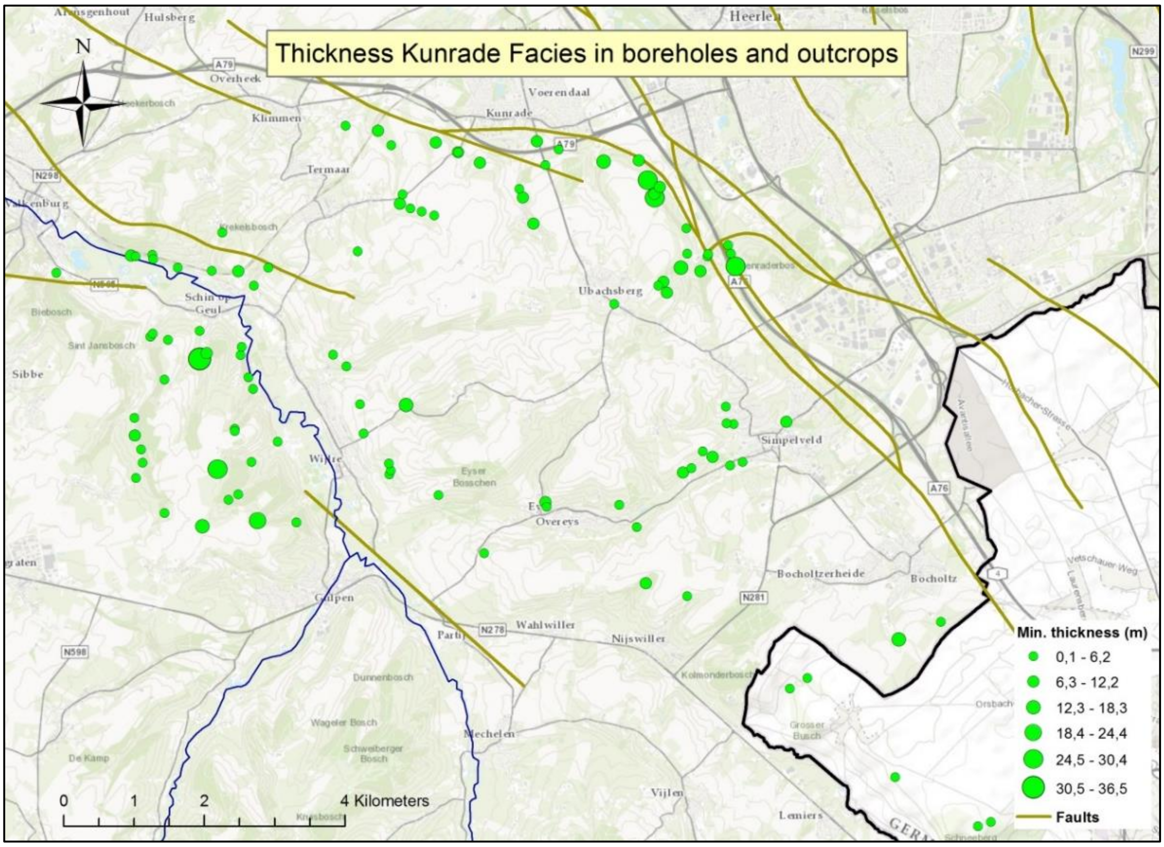

Fig. 4.10.3. Thickness of the Kunrade Facies. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Fossil content
The sediments are characterized by specific benthic foraminifera, grouped in Zones G, J and O (Hofker, 1966). Also distinguished as the 'sponges facies' by Hofker (1966) due to the large abundance of sponge-spicules in the sediments (Fig. 4.9.12). The Kunrader Facies is the equivalent of the Maastricht Facies in the west (Hallembaye and Maastricht) and the transitional Schaelsberg Facies in the middle of the study area (Valkenburg and Cadier en Keer) as proposed by Felder (1975). Felder et al. (1985 and 1989) showed by the study of bioclasts and ostracodes that this is more or less correct as the fossil content is equal to the lower half of Zone J (sensu Hofker, 1966) in Valkenburg and Zone H (sensu Hofker, 1966) in Maastricht. However, Felder et al. (1985 and 1989) also suggested that a part of the Lanaye Member is the equivalent of the lower part of the Kunrader Facies based on biostratigraphy.

The fossil content is dominated by remains of echinoderms (crinoids) and bryozoans, also clasts of molluscs are common (Felder et al., 1989). The combined number of clasts from bryozoans, corals and sponges is high and ranges between 0 and 70%, commonly it reaches over 30% of the total fossil content. Felder et al. (1989) states that the lower part of this facies unit can easily be distinguished by frequent high numbers of crinoids clasts (up to 22%) and the common presence of octocorallia rods. In the upper part the sediments are characterized by relatively low numbers of crinoid clasts (rarely exceeding 5% of the total bioclasts) and lower numbers of octocorallia rods. The number of serpulid clasts is only very small and is slightly increasing towards the top of this facies, with the *Sclerostyla macropus* as characteristic species (Felder et al., 1989). This facies unit is characterized by a relatively high number of ornamented ostracods.

Age
The age of the Kunrade Facies is Late Maastrichtian based on belemnite zonation (Schmid, 1959; Schulz et al., 1984).

Interpretation of depositional environment
The grain size of bioclasts in these calcarenites is coarser than stratigraphically older sediments which indicates deposition in a relatively higher energetic environment. The relatively high content of siliciclastic grains indicates that this area received input from zones of erosion, so a relatively proximal environment.

The alternation of hard and soft beds is caused by a difference in cementation. Cementation in marine sediments is very common in shallow marine environments and can be related with regular wave activity during deposition (Indin et al., 1983). Zijlstra (1995) also showed that there is a positive relation between the degree of cementation and hydrodynamic energy of the depositional environment. The change from dominantly hard cemented beds in the east to dominantly soft and less cemented beds in the west, can be interpreted as a decrease in energy conditions during deposition.

The relative abundance of bryozoans, corals and sponges is a lot higher in the Kunrade Facies than it is in the Maastricht Facies which indicates relative higher hydrodynamic energy conditions. This is confirmed by the relatively large number of ornamented ostracods. Felder (1985) states that ornamented ostracods are linked with a relatively higher energetic environment than soft-shelled ostracods. The lower part of the Kunrade Facies contains relatively more crinoids than the middle and upper part. This indicates that the lower part was deposited in a depositional environment with lower hydrodynamic energy conditions. The decrease in crinoids in the middle and upper part indicates an increase in hydrodynamic energy conditions, that can be interpreted as a relative shallowing of the depositional environment. Towards the west (Maastricht Facies), the number of crinoids are increasing indicating a decrease in hydrodynamic energy conditions, interpreted as a deepening of the depositional environment.

The serpulid species *Sclerostyla macropus* is typical for the Kunrade Facies and is absent in the Maastricht Facies where *Sclerostyla regia* forms the characteristic species (Felder et al., 1989). Jäger (1988) stated that *Sclerostyla macropus* is characteristic for boreal shallow-marine environments, this means that there was less influence from warm oceanic waters from the Tethys Ocean. Since this species is not found in the Maastricht Facies it can be suggested that the Maastricht Facies is relatively deeper than the Kunrader Facies. It is also likely that boreal conditions were less dominant in the Maastricht Facies, while the warm oceanic water of the Tethys Ocean were of greater importance.

4.11 Nekum Facies

The Nekum Member (Felder, 1975) can be regarded as a specific facies, called the Nekum Facies. It is characterized by calcarenites with flint nodules in the lower part. According to Bless et al. (1987) the grain size of bioclasts is 120 to 150 µm. Following the Dunham classification the texture of these deposits can be classified as packstones and grainstones. A facies subdivision has been made: Nekum Facies A and Nekum Facies B.

The Nekum Member is recognized in boreholes and outcrops in the northwest of the study area (Fig. 4.11.1). Towards the Roer Valley Graben the Nekum Facies has not been recognized due to subsidence of the Roer Valley Graben. The Cretaceous strata are covered by a thick succession of younger deposits and hence difficult to reach. The thickness of the Nekum Member varies from 9 to 14.4 m, on average a thickness of 10 to12 m is found (Fig. 4.11.2). The thicknesses increases towards the west.

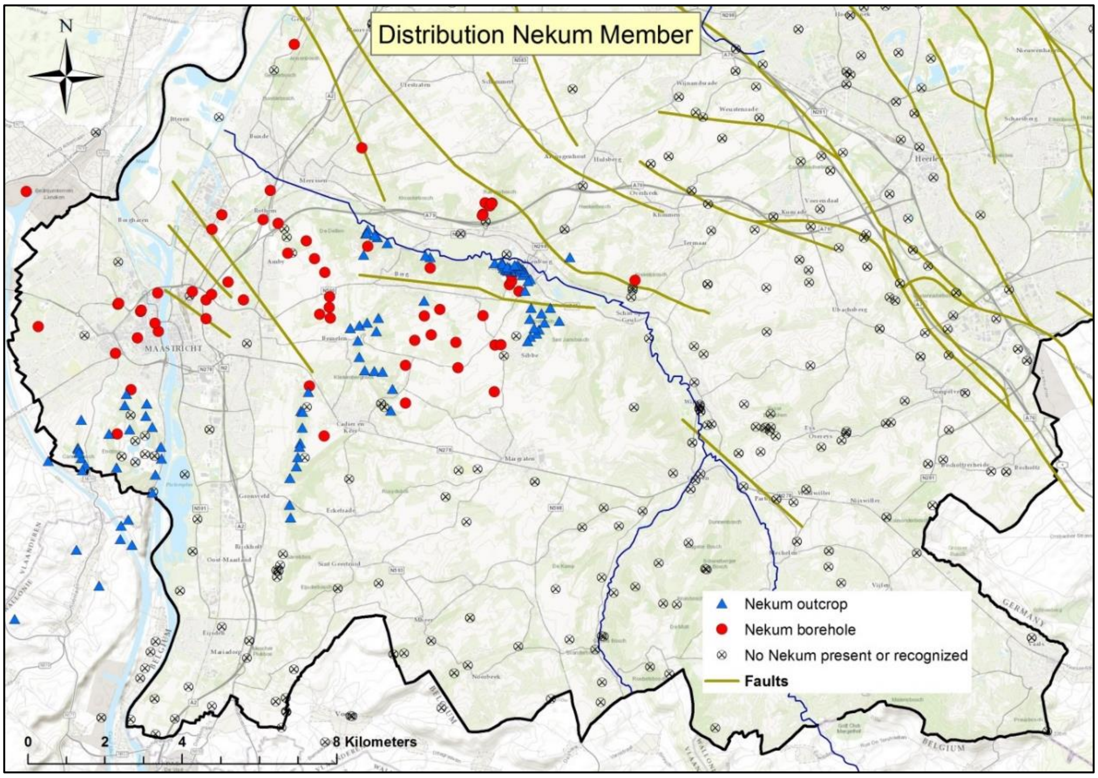

Fig. 4.11.1. Distribution of locations where the Nekum Member is found.

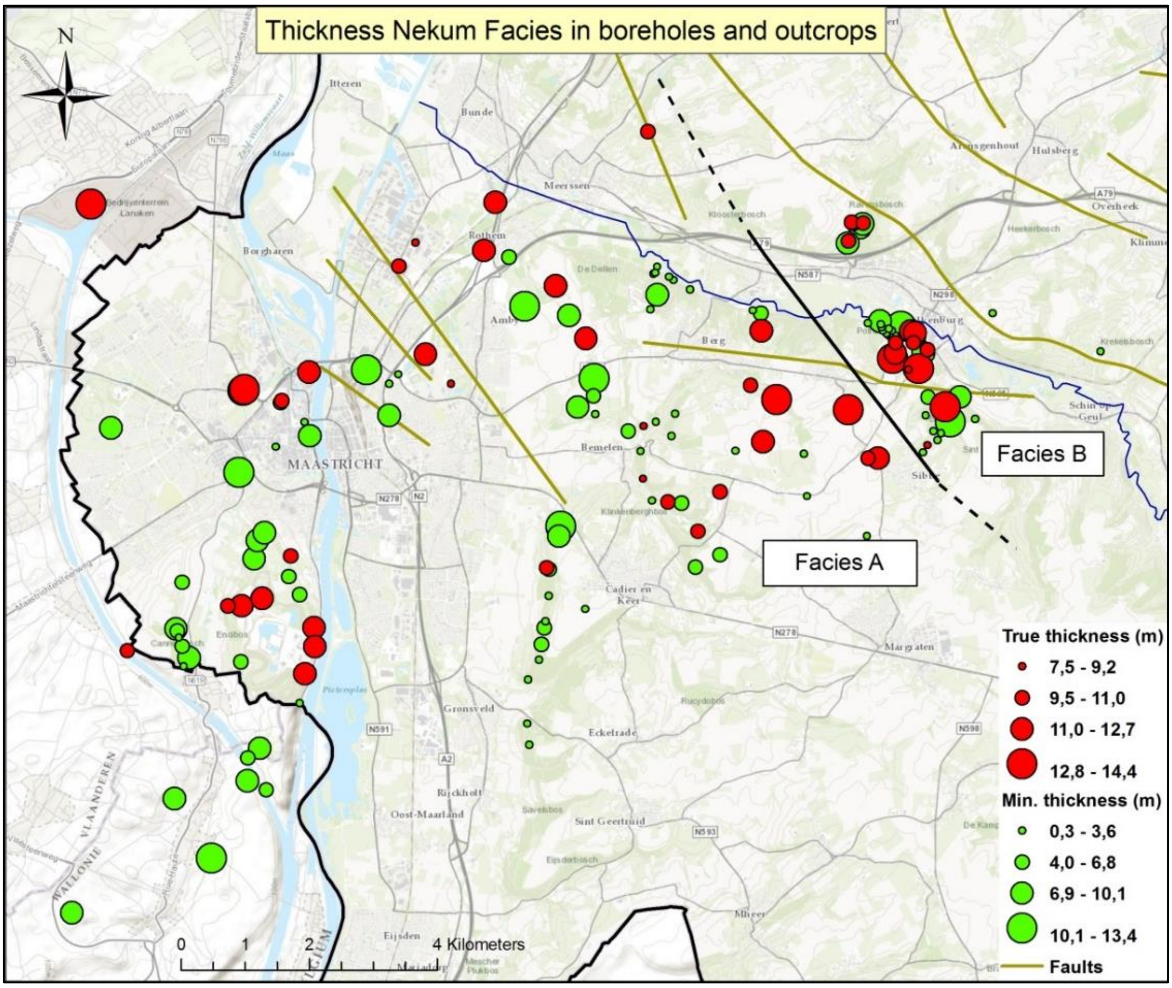

Fig. 4.11.2. Thickness of both Nekum facies. The true thickness is the thickness when Cretaceous deposits are found on top of the Nekum Member. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments). NE of the black colored line Facies B is found, whereas toward the SW Facies A is present.

Within the Nekum Member two facies can be identified. They laterally grade into each other.

Nekum Facies A

Nekum Facies unit A is a homogeneously bioturbated calcarenite, with at the base a few decimetre thick coarse-grained fossil-grit layer that is laterally continuous throughout the study area. This basal grit-layer covers a hardground with a wavy top surface, called the Horizon of Laumont (Fig. 4.11.3). In some locations the coarse-grained base of this facies shows small-scale cross-bedding and lamination (Zijlstra, 1995). Three to six metres above the basal fossil-grit layer another lateral continuous fossil-grit layer is distinguished on top of a slightly indurated layer. In between, flint nodules have formed during diagenesis. The upper fossil-grit layer is called the Horizon of Kanne by Felder (1975), it can be tracked throughout the study area (Fig. 4.11.4). Above this horizon, deposition of calcarenites continued, alternated with few centimetre thin fossil-grit lenses that are laterally continuous for only several metres (Fig. 4.11.5). Villain (1977) recognized centimetre-scale crossbedding in these fossil-grit layers. In the top of this member a hardground formed, named the Horizon of Caster (Felder, 1975). Villain (1977) and Zijlstra (1995) recognized a fining upward grain size.

Fig. 4.11.3. The picture shows the underlying Emael Member (below) and the Nekum Member (above) in the quarry Sibbergroeve south of Valkenburg. The contact between these two units consists of a hardground (HG), which contains a wavy top surface (black arrows). Above this surface, the basal fossil-grit layer which forms the base of the Nekum facies unit. The vertical distance is approximately 2 to 2.5 metres. (Photo: Jac Diederen)

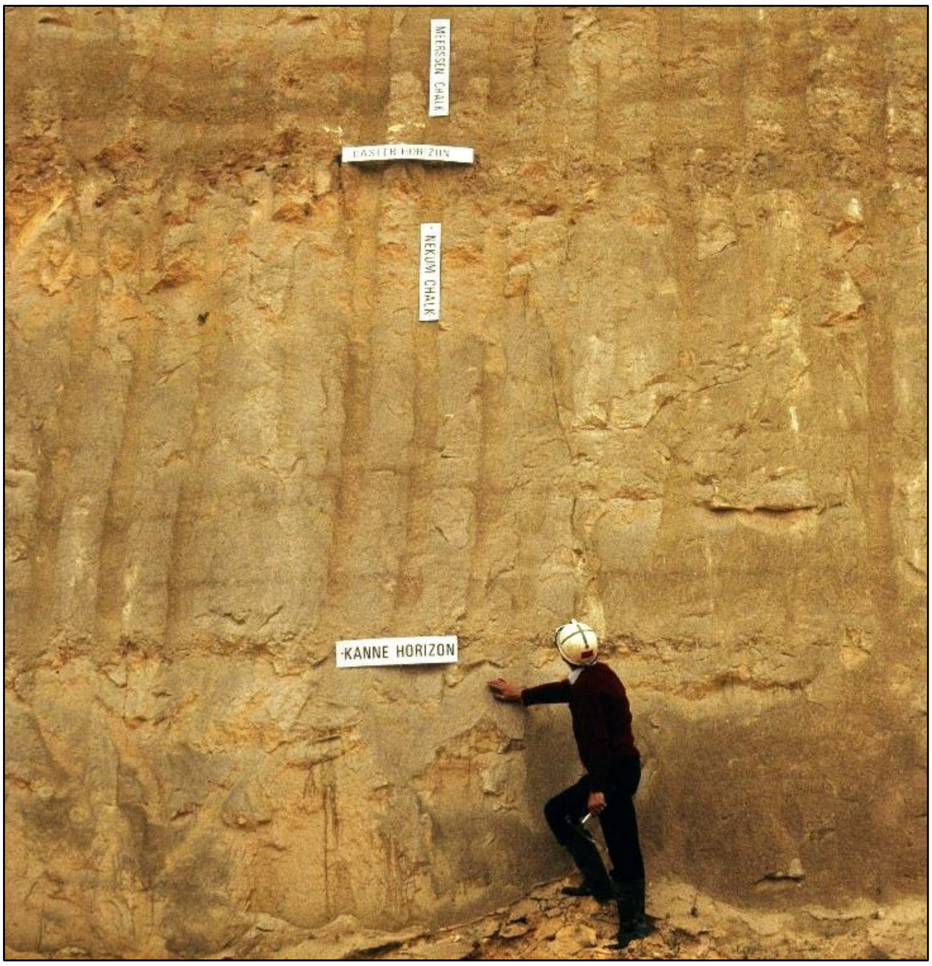

Fig. 4.11.4. Sediments of the Nekum Member with the Horizon of Kanne and the Horizon of Laumont in the top, covered by sediments of the Meerssen Member. Quarry ENCI. (RGD)

Measurements of the $CaCO_3$ content in the quarry ENCI, show that the Nekum Member contains an average of 98% carbonate content measured on four different locations (TNO – RGD, report OP5730-3). The remaining few percent are clay or other clastic particles. At the base of the facies unit a small carbonate dip can be observed (96%), while the largest part shows a percentage of 98 to 99%. Quarry ‘t Rooth, 6 km to the east of Maastricht, shows the same value: 98.5% to 99% (TNO - GB, report GB790). Quarry Curfs, west of Valkenburg (62A-13) shows an average of 96.7% for the upper part. It appears that the average carbonate content decreases towards the northeast and that the base of this facies unit contains a slightly lower carbonate content than the upper part. Nekum Facies A is most common and all sediments south and west of the village of Valkenburg are classified as Nekum Facies A (Fig 4.11.2).

Nekum Facies B

Nekum Facies B is a calcarenite with alternating hard and soft discontinuous beds and fossil debris lenses on top of a few decimetre thick basal fossil-grit layer. The basal grit-layer covers a hardground with a wavy top surface (Horizon of Laumont), and is identical to the basal layer in Nekum Facies A (Fig. 4.11.3). In some locations the coarse-grained base of this facies shows small-scale cross-bedding and lamination (Zijlstra, 1995). The second fossil-grit layer that is laterally continuous, is the same as described in Nekum Facies A. However, below the second fossil-grit layer a cemented layer of 10 to 20 cm thick forms a proto-hardground, this has been called the Horizon of Valkenburg (Felder, 1975). In between the basal fossil-grit layer and the Horizon of Valkenburg flint nodules are found but in lower quantities than Nekum Facies A (Fig. 4.11.5).

Above the Horizon of Valkenburg, calcarenite deposition continued. The sediments show an alternation of discontinuous hard and soft beds as well as more developed centimetre-thick fossil-grit lenses than found in Nekum Facies A (Fig. 4.11.5). In the top of this facies a hardground has formed, which is the same hardground as in Nekum Facies A (Horizon of Caster). Villain (1977) and Zijlstra (1995) recognized a fining upwards in grain size.

Outcrops where Nekum Facies B can be recognized are located in and around the village of Valkenburg (Fig. 4.11. 2). Table 4.11.1 shows the most important outcrops where this facies is found. Since its distribution is limited to a small area without large industrial quarries, no $CaCO_3$ content measurements are known.

Identification number of location (TNO/RGD)	Name of location (outcrop or quarry)
62A30	Quarry Biebosch
62A215	Gemeentegrot
62A289	Lemmekeskoel
62A197	Flessenberg
62A401	Dwingel castle
62A518	Openluchttheater
62A519	Fluwelengrot

Table 4.11.1. Names and identification numbers (according to TNO/RGD) of the most important outcrops where Nekum Facies B can be observed.

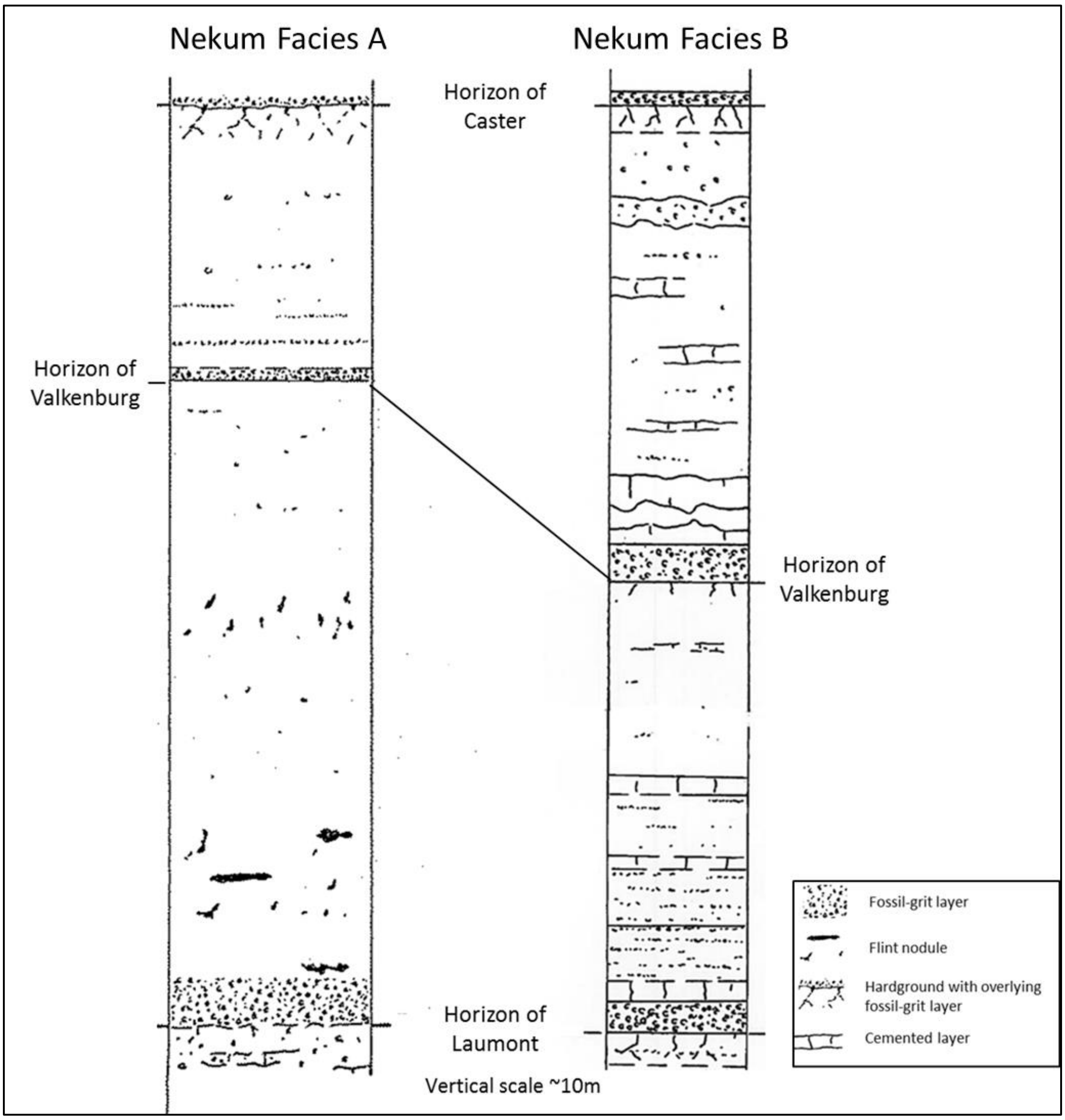

Fig. 4.11.5. Sketches modified after W.M. Felder of outcrop 61F019 (Nekum Facies A) and 62A402 (Nekum Facies B). (TNO/RGD outcrop database)

Age

The age of the Nekum Member is Late Maastrichtian (Felder et al., 2000).

Fossil content

Remains of the sea urchin *Hemipneustes* are very common in the Nekum Member. The highest concentration is reached in a layer called the '*Hemipneustes* level', described by Felder (1963). Fossil content in the coarse-grained base is dominated by broken clasts of *Pycnodonte vesicularis* (oysters*), Sclerostyla mosae* (serpulids), bivalves, gastropods and rare ammonites (Villain, 1977; Felder et al., 2000).

This is confirmed by Felder (2001) who stated that the bottom metres of this facies are characterized by a large number of serpulids. According to Jäger (1987), serpulids become more dominant above the base of the Emael Member and in the Nekum Member of Felder (1975). The presence of serpulids is indicative for a warm and shallow environment, and suggests the presence of a solid substrate (Felder, 2001). *S. Mosae* is most common, this species preferred relatively quiet water conditions in comparison to other serpulid species (Jäger, 1987). In the stratigraphically higher Meerssen Member, *S. Mosae* is less common, and hence the environment was probably more energetic. The genus *Cementula* becomes more common in the Nekum Member. According to Jäger (1987) this genus is indicative for warm and shallow waters. Similar species are currently found in the Caribbean in water depths ranging from 13 to 100 m (Hove, 1973). The high abundance of serpulids is accompanied by a high abundance of bryozoans (Felder, 2001). Also bivalves are found in sediments, their abundance is not as high as in sediments of the Gulpen Formation but do show a distinct peak at the base of this member. These sediments contains the highest abundance of molluscs (oysters) in the Maastricht Formation, and can be recognized almost everywhere in the study area. The top of these sediments also show a small number of crinoids (Felder, 2001).

Both Nekum Facies are difficult to distinguish in borehole data, especially at great depth close to the Roer Valley Graben. However. Some typical fossils for the Nekum Facies are recognized in boreholes, which indicates that it is presumably present on the southwest rim of the Roer Valley Graben. Felder (TNO - GB, GB1917) recognized specific species of the Nekum facies in borehole B60C0897 (Sittard). The lithology is briefly described as limestone with some glauconite. The fossil content is dominated by an increase in foraminifera, molluscs and brachiopods relative to sediments in the southwest. Numbers of bryozoans decreased and the number of echinoids remained roughly constant. Remains of arthropods are regularly present in deposits in Maastricht and Valkenburg, mainly remains of *Callianossa* (a genus of mud shrimps). However, in borehole B60C0897 at the southwest rim of the Roer Valley Graben this genus disappears out of the fossil assemblage. Presumably conditions were less favourable towards the Roer Valley Graben, which is likely caused by a lack of soft substrate needed for *Callianossa.*

Interpretation of depositional environment

Nekum Facies unit A is a homogeneously bioturbated calcarenite with thin fossil-grit layers, while Nekum Facies unit B is more heterogeneous and shows an alternation of hard and soft beds with more pronounced fossil-grit layers (Fig. 4.11.3). The harder beds are more cemented, which is very common in shallow marine environments that are subject to regular wave activity (Inden et al., 1983). Cemented layers have formed identical to hardgrounds, with the exception that the degree of cementation is less pronounced (Felder, 1973). Zijlstra (1995) shows that there is a positive relation between the degree of cementation and hydrodynamic energy. This suggests that the lithified layers in Nekum Facies unit B were deposited in an environment with higher hydrodynamic energy conditions, hence a shallower depositional environment than deposits of Nekum Facies A. This is in agreement with the disappearance of the mud shrimp *Callianossa* towards the northeast due to the lack of soft and muddy substrates. Also the decrease of carbonate content, and hence the increase of siliciclastic grains, points to more proximal conditions towards the northeast. Whereas in the southwest (Nekum Facies A), energetic conditions were relatively lower as demonstrated by the lack of cemented layers and only thin fossil-grit layers. Also the presence of *Callianossa* points to a more distal marine environment with a soft and muddier sea floor.

The common basal fossil-grit layer on top of a hardground in both facies units, points to similar conditions during deposition. Zijlstra (1995) interpreted this as a tempestite, a storm deposit. The abnormally high numbers of serpulids and oysters suggests the presence of a solid substrate which could have been the cemented hardground below that is partly eroded. This is evidenced by the wavy top of the hardground which suggests erosion by wave activity. Zijlstra (1995) suggested that the formation of the hardground occurred during a period of increased storm intensity. Zijlstra (1995) proposed a Milankovitsch-related cyclicity in storm intensity. At maximum storm intensity the sediments would be repeatedly eroded and transported elsewhere, causing a period in which no new sediments were deposited. When storm intensity decreased, storm waves reworked the sediment but no significant transport occurred. This process caused the formation of crossbedding and lamination (tempestite formation), as well as the accumulation of eroded fossil remains. Only occasionally these sedimentary structures are preserved, most of the time these were destroyed by burrowers. As storm intensity kept decreasing, the hydrodynamic conditions decreased and the rate of sedimentation slowly increased as less sediment is removed during storms. Occasionally, sedimentation of calcarenites is disturbed by thin lenses of coarse-grained fossil-grit layers, deposited by storms.

The upper continuous fossil-grit layer (Horizon of Kanne and Horizon of Valkenburg) formed by the same process. The genesis is similar to the genesis of the basal-grit layer. Possibly, the depositional environment deepened or was positioned in a more quiet depositional setting, resulting in lower energy conditions during maximum storm intensity as shown by the absence of a hardground in the west of the study area. This hypothesis is backed by the observation that the grain size is fining upwards towards the Nekum Member. The presence of a proto-hardground underneath the Horizon of Valkenburg (Nekum Facies B) in the area of Valkenburg, suggests deposition in a higher energetically environment than the Horizon of Kanne (Nekum Facies A).

The Nekum Member is described as a calcarenite which formed by winnowing of the sediment and sorting of different grain sizes. This indicates that the sediment surface was in reach of waves, so a depositional environment above the fair wave base is proposed. The low amount of clastic grains suggests some distance from the shoreline and tectonic quietness.

4.12 Meerssen Facies

The Meerssen Member can be classified as a specific facies, called the Meerssen Facies. It is characterized by calcarenites with hardgrounds and fossil-grit layers. The mean grain size of bioclasts ranges from 120 to 150 µm (Bless et al., 1987). According to the Dunham classification the structure of these sediments can be classified as packstones and grainstones. Its base is formed by the Horizon of Caster that forms the transition from the top hardground in the Nekum Member and the basal fossil-grit layer of the Meerssen Member (Felder, 1975). Its top is formed by the Horizon of Vroenhoven which is the top of a hardground. The hardgrounds make up 10% of the total thickness (Felder et al., 2000). Fining-upwards cycles can be observed in between hardgrounds, with the coarsest grains in the fossil-grit layer covering the hardground. The fossil-grit layers reach a thickness up to 2 m. Flint is absent. The thickness of these cycles ranges from 0.5 m to multiple metres. The thickness of individual cycles increases towards the top of the Meerssen Member (Felder et al., 2000). The $CaCO_3$ content ranges between 95 and 99%, and south of Maastricht it reaches 98% on average (ENCI quarry).

The distribution of the Meerssen Member is limited to the northwest of the line Gronsveld – Valkenburg (Fig. 4.12.1). In the area of Valkenburg, Maastricht and Meerssen this member crops out, further towards the northwest its presence is confirmed in all boreholes that reach sufficient depth. Some boreholes in the Roer Valley Graben also contain sediments of this member, confirmed by biostratigraphy. One of these, B60C-897 in Sittard, shows a thickness of 9.5 m for the Meerssen Member. The Meerssen Member reaches a thickness of 18 m on average (Fig. 4.12.2).

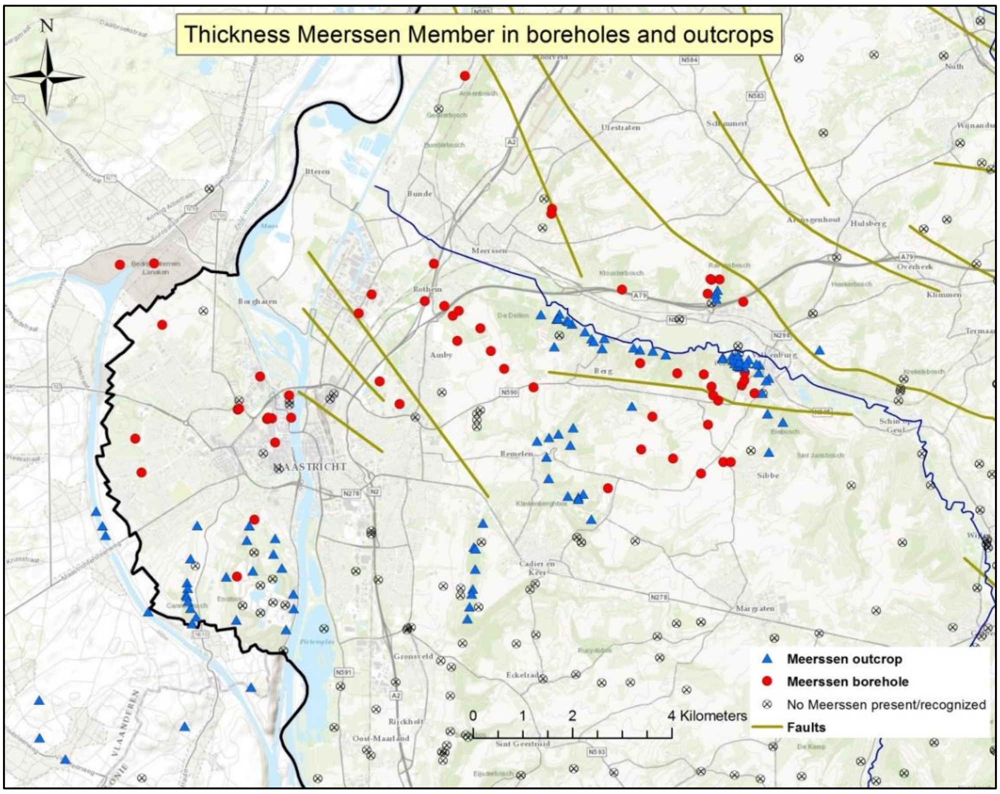

Fig. 4.12.1. Distribution of locations where the Meerssen Member is found.

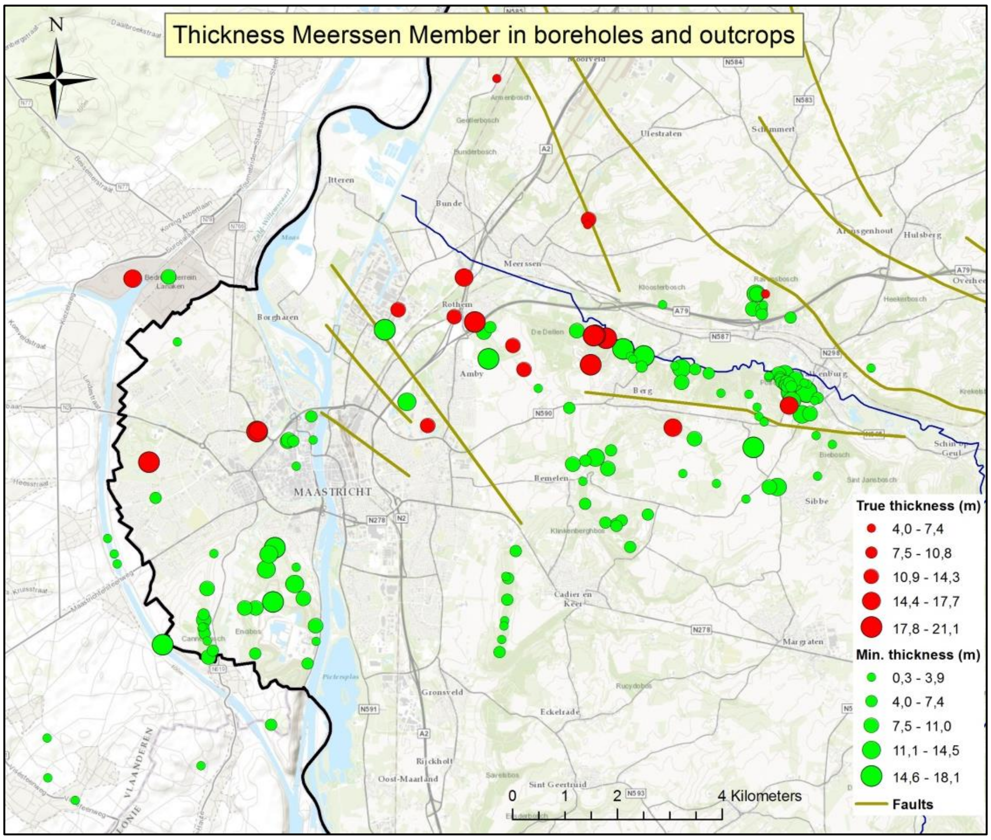

Fig. 4.12.2. Thickness of the Meerssen Member. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Fossil content
Hofker (1966) created three zones of benthic foraminifera assemblages, from base to top: zone L, M and N. Zone N was only occasionally found, and later appeared to be of Danian age (Herngreen et al., 1998).

The fossil-grit layers contain many remains of bryozoans, corals, large foraminifera species and anthozoans. Herngreen et al. (1998) recognized two different palynomorph assemblages in the Meerssen Member (Quarry Curfs). From the top of the Nekum Member to a few metres above the Horizon of Caster, the palynomorph assemblages are relatively poor and dominated by *Paralecaniella spp*. From a few metres above the Horizon of Caster into the Houthem Formation (Danian) the palynomorph assemblages are relatively rich and dominated by dinoflagellate cysts.

Borehole B60C-897 (Sittard) in the Roer Valley Graben has been extensively studied on fossil content (TNO - GB, GB1917). It appeared that the top of the Meerssen Member (Zone N) is not present. The base of the Meerssen Member shows a relatively large number of bryozoans in quarry Curfs and in Borehole B60C-997 (Sittard). Towards the northwest (The Belgium part of the Campine Basin), borehole KS18 lacks these large numbers of bryozoans and echinoids are more dominant.

Age
The age of the Meerssen Member is Late Maastrichtian and its sediments are the last Cretaceous sediments deposited. However, between Valkenburg and Meerssen (Geulhemmergroeve, and Quarry Curfs, 62A-13) it appeared that the Horizon of Vroenhoven was already deposited in the Paleocene (Brinkhuis et al., 1996; Herngreen et al., 1998). The top of the Cretaceous is found up to 3.5 m lower than expected, present as an interval of clay (Horizon of Berg en Terblijt). Moving away from this area the vertical distance of the two horizons decreases and eventually the Horizon of Berg en Terblijt disappears and the Horizon of Vroenhoven forms the top of the Cretaceous sediments (Felder et al., 2000)

Interpretation of the depositional environment
The coarse grain size of bioclasts indicates that deposition occurred in a relatively high-energy environment, more energetic than the previous discussed facies. This is also shown by the abundance of coarse-grained fossil-grit layers which also indicates that storm deposits are common. The abundance of hardgrounds indicates that periods of non-deposition were common as well. This however, seems contradictory. The genesis of these hardgrounds can be explained by accepting a cyclicity in storm intensity, as is explained in Chapter 3.

During an increase of storm intensity and related hydrodynamic energy, the increased erosion rates equalled the subsidence rates and the deposition rate was consequently zero resulting in a lithified layer (Zijlstra, 1995). During an increase in storm intensity, the sediment that had been lithified below the sediment surface, was repeatedly eroded, exhumed and exposed during storms (Zijlstra, 1995). During maximum storm intensity, the lithified sediment was continuously exposed and a a bored and encrusted hardground formed.

Villain (1977) interpreted the depositional environment as very shallow marine, with water depths of 2 to 15 m. The presence of solitary corals indicates that paleo-water temperatures reached 20 to 25 degrees Celcius (Liebau, 1978). Herngreen et al. (1998) interpreted the depositional environment of the top of the Meerssen Member as inner neritic with depths of less than 30 m, characterized by benthic foraminiferal faunas while planktonic foraminifera and ostracods are rare.

The base of the Meerssen Member is dominated by *Paralecaniella spp.* which is indicative for extremely marginal marine environments, with relatively higher energy conditions than the zone dominated by dinoflagellate cysts (Brinkhuis et al., 1996). The middle and upper part of the Meerssen Member is dominated by dinoflagellate cysts, which reflect outer neritic conditions and relatively lower energy conditions during deposition (Herngreen et al., 1998). From the base of the Meerssen Member upwards, the general trend can be interpreted as a relative deepening of the depositional environment. In detail, a cyclicity can be observed in *Paralecaniella spp*- dominated to dinoflagellate cysts-dominated environments. These appear to be influenced by third-order sea level changes, since the timespan of the Meerssen Member is in the order of 2 million years (Haq et al., 1988; Herngreen et al., 1998). Hengreen et al. (1998) allocated system tracts to different intervals. Intervals dominated by dinoflagellate cysts are presumed to represent a Highstand Systems Tract (HST), whereas Lowstand Systems Tract (LST) are missing but are represented by hardgrounds, followed by a Transgressive Systems Tract (TST) that shows a gradual change from *Paralecaniella spp*- dominated to dinoflagellate cysts-dominated assemblages. The TST ends with a Maximum Flooding Surface, represented by a peak in the number of dinoflagellate cysts. This is again followed by a HST (Herngreen et al., 1998). The presence of many more hardgrounds that are less well developed than the Horizon of Caster and Horizon of Vroenhoven, may represent LST's caused by fourth-order sea level changes. The Horizon of Berg en Terblijt, that represents the transition from Cretaceous to Paleogene was deposited during a HST, followed by a LST (Horizon of Vroenhoven).

The southern extent of the Meerssen Member is limited by the line Gronsveld - Valkenburg. This is caused by the Cenozoic uplift of the Ardennes that commenced in the southeast (Demoulin, 1995). During deposition this uplift was absent and it is likely that the Meerssen Member was deposited over a larger area. In the Roer Valley Graben, the Meerssen Member is less thick and represented by relatively large numbers of bryozoans. Bryozoans prefer shallow marine conditions, a hard substrate, as well as relatively high energy conditions (Felder et al., 2000). This may suggest that the Roer Valley Graben was still inverted and subsidence was occurring only slowly. This is confirmed by the reduced thickness of the Meerssen Member and the absence of the upper part of the Meerssen Member.

4.13 Houthem Formation

After the Cretaceous, sedimentation of carbonates continued into the Danian with the Houthem Formation (Herngreen et al., 1998). The Houthem Formation is subdivided in three members: Kalksteen van Geulhem, Kalksteen van Bunde and Kalksteen van Geleen (Felder, 1975). Only the Kalksteen van Geulhem is exposed and well described. The Kalksteen van Bunde and the Kalksteen van Geleen are only known from boreholes and lithology is only scarcely described. It appears that the lithology is roughly similar for all members, so the Houthem Formation will be discussed as one unit. The sediments are characterized by calcarenites with hardgrounds and fossil-grit layers that show a variation in thickness. The sediments show an alternation of layers that contain hard nodules or lenses and layers that are relatively soft. Glauconitic grains are present in some beds (Felder et al., 2000). The $CaCO_3$ content ranges from 96 to 97% in the Kalksteen van Geulhem, and from 98 to 99% in the Kalksteen van Bunde (Felder et al., 2000)

The Houthem Formation is only present in the northwest part of the study area (Fig. 4.13.1). To the southeast older strata are exposed. The thickness ranges from 0 to 80 m, the greatest thicknesses are reached in the northwest of the study area, north of Maastricht (Fig. 4.13.2). North of the

Heerlerheide Fault the thickness decreases. The Houthem Formation is absent north of the Feldbiss Fault.

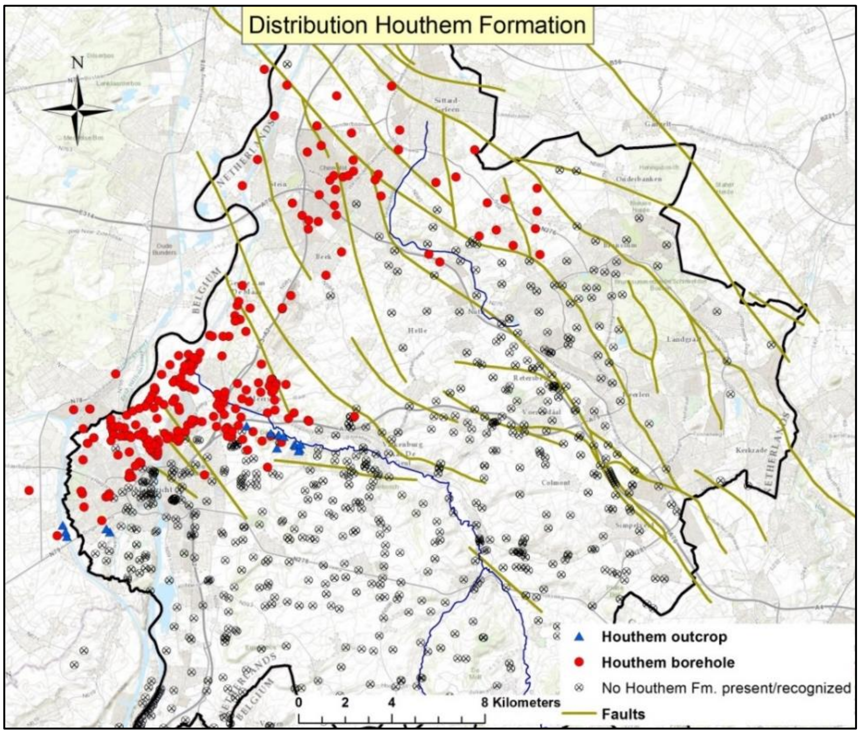

Fig.4.13.1. Distribution of locations where the Houthem Formation is found.

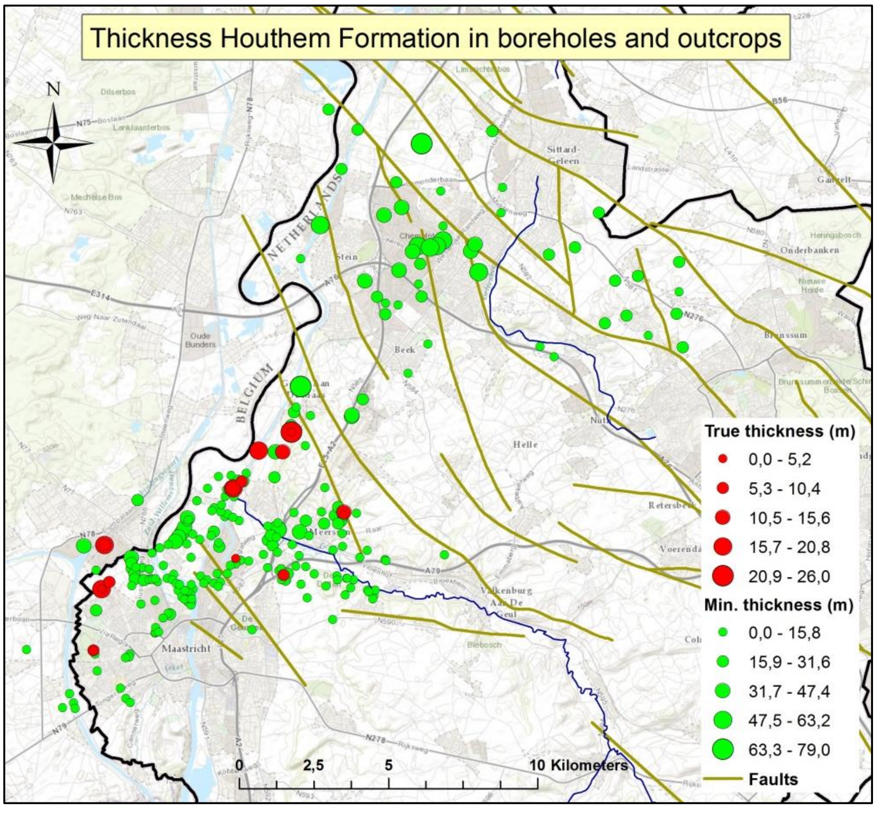

Fig. 4.13.2. Thickness of the Houthem Formation. The true thickness is the thickness when Cretaceous deposits are found on top of the sediments of this unit. The minimum thickness is the thickness obtained from locations where the bottom is not exposed or the top is not exposed or eroded and covered by Cenozoic deposits (Oligocene or Pleistocene sediments).

Interpretation of depositional environment

The sediments show an alternation of layers that are relatively soft and layers with relatively hard nodules. This can be interpreted as a difference in degree of cementation, which is common in depositional environments subject to wave activity (Inden et al., 1983). The presence of hardgrounds and fossil-grit layers indicate that episodic phases of non-deposition followed by storm deposits occurred (Zijlstra, 1995). This is however, less clear than in the Meerssen Member. This may indicate a slightly deeper depositional environment as many layers are cemented but not well enough to form a hardground. This is confirmed by research of Herngreen et al. (1998), who indicated that the lower part of the Houthem Formation was dominated by dinoflagellate cysts. The domination of dinoflagellate cysts was interpreted to reflect outer neritic conditions and relatively lower energy conditions during deposition (Herngreen et al., 1998). The relative deepening of the depositional environment can be explained by an increase in relaxation of the inverted Roer Valley Graben resulting in increased subsidence. The thickness map shows that this is only partly true, because the thickness of the Houthem Formation decreases north of the Heerlerheide Fault. The thickest succession is found in the northwest part of the study area which indicated that accommodation space was greatest here. This also indicates that the compressional regime was still influencing sedimentation in the Roer Valley Graben.

5. Discussion

5.1 Lithostratigraphic and biostratigraphic subdivisions and facies

The classic lithostratigraphic subdivision of Felder (1975) has been used as a framework to study changes in lithology and to come up with a facies analysis. This subdivision was matched with a biostratigraphic subdivision based on benthic foraminifera (Hofker, 1966) and with a chronostratigraphic subdivision based on belemnites (Schmid, 1959). Benthic foraminifera can be used for a facies analysis as different species reflect different depositional environments. If a lithostratigraphic and biostratigraphic subdivision can be matched, than the depositional environment per lithostratigraphic unit must have been quite uniform. However, some gradual changes in lithology can be observed laterally per lithostratigraphic unit. A facies analysis is done in this study to reveal changes in depositional environment per unit.

Most lithostratigraphic units can be interpreted as individual lithofacies as the lithology remains the same, and hence also the depositional environment to some extent. The depositional environment of the Maastricht Formation was not similar everywhere, as is shown by the Kunrader and Maastricht Facies. The Kunrader Facies has been regarded as the coeval equivalent of the Maastricht Facies since the 19th century. A satisfying explanation based on lithological correlation has never been made. The missing link appeared to lie in correlation based on biostratigraphy (Robaszynski et al., 1985; Felder et al., 1985, 1989). Felder et al. (1985) suggested that the upper part of the Lanaye Member, as well as the lower part of the Maastricht Formation (Valkenburg, Gronsveld, Schiepersberg and Emael Member) was the equivalent of the Kunrade Facies.

This would also explain the great differences in thickness of the Valkenburg Member, as well as the complete change in lithology from west to east. As the fossil content, coarsening upwards and sedimentary structures show, both facies were deposited during different environmental conditions. This correlation could not have been created based on lithology, correlation based on biostratigraphy appeared to be useful. Felder (1975) accepted the idea that the lithology can change laterally. On formation scale however, he did not, and assumed that the base and top of a lithological formation are coeval in different locations. This implies that for example the base of the Maastricht Formation (Horizon of Lichtenberg) is not necessarily of the same age in different locations. It is important to realize that the subdivision of Felder (1975) is based on lithostratigraphy, which does not incorporate timelines and hence cannot be used for chronostratigraphy. In other words, the classical idea that the Gulpen Formation is entirely older than the Maastricht Formation is false. In a specific period of time it is logical that a depositional environment closer to shore and more shallow marine corresponds with other sediments than a depositional environment more distal and deeper marine. By making a facies analysis new insights and additional information about the depositional environment is obtained.

5.2 Sediment controls

The largest part of the sediments consist of biodetrital calcisiltites and calcarenites, as well as a a varying amount of siliciclastics. In several large quarries, the $CaCO_3$ content is measured which indicates the relative amount of carbonate material. The remaining siliciclastic material is dominantly SiO_2 and small percentages of Fe_2O_3. The calciclastic and siliciclastic material is of allogenic origin and has been transported and redeposited. The biodetrital carbonates are likely to originate from the basin. Siliciclastic grains were eroded and transported and originate from areas where non-carbonate sediments were eroded. This may have been in a terrestrial setting, in which siliciclastic sediments where transported into the basin by river plumes. Another possibility is erosion of non-carbonate sediments inside the basin. Both profiles (Appendix) show large thickness variations for the Vaals and Aken Formation which are both characterized by fine-grained sand. I believe that these variations in thickness is caused by preferential uplift along different faults during inversion of the Roer Valley Graben. This would indicate that the source of siliciclastic sediments is quite local as is shown by unexpected high numbers of siliciclastic grains in biodetrital sediments along faults (borehole B62A0309). If we accept that the Roer Valley Graben is the main source of siliciclastic sediment than the relative amount of siliciclastic grains can be used as a rough indication of proximity to the Roer Valley Graben. This is confirmed by the observation that intraformational clasts, bioclasts and siliciclastic grains increase in size towards the Roer Valley Graben. In the northeast of the study area, Carboniferous pebbles and cobbles are found indicating erosion of Carboniferous strata in the inverted Roer Valley Graben. These are often found in layers with coarse-grained siliciclastic sediments, the clasts increase in size towards the Roer Valley Graben. It is difficult, if not impossible, to distinguish between autogenic and allogenic controls. In general, the amount of transported siliciclastic grains decreases with increasing distance from the source area as the hydrodynamic energy conditions decrease. On the other hand, increased erosion rates may be caused by a temporal increase in uplift of the Roer Valley Graben, which resulted in increased transport of siliciclastic grains towards the basin.

Most of the sediments contain glauconite in varying amounts. This may have formed in-situ but can also be allogenic. The Vaals Formation contains the highest amount of glauconite and may have acted as source during erosion of the inverted Roer Valley Graben.

5.3 Water depth reconstruction

To come up with a water depth reconstruction the lithology, texture, flint nodules and sedimentary structures are used.

Distribution of flint nodules can be used as an indication for the paleo-water depth. As explained in Chapter 3.1, late diagenetic precipitation of flint takes place where there already is an increased concentration of early diagenetic silica polymorphs (Zijlstra, 1995). Since the early diagenetic silica concentration was inversely proportional to the deposition rate, an increase in size of flint nodules indicates no or decreased sedimentation. This implies that continuous flint nodule layers formed during relative low energy conditions, as well as a period of non-deposition. Randomly distributed flint nodules formed during high energy conditions as the sediment surface was occasionally reworked by storms. The thickness of these flint nodules are also less than the nodules in flint nodule layers, which confirms the relative high energy conditions that cause a shorter residence time in the zone of silica precipitation during diagenesis.

Fossil-grit layers are common in the Maastricht Formation and at the base of members of the Gulpen Formation. The presence of fossil-grit layers, alternated with biodetrital chalk, can be interpreted as caused by episodic flows. Episodic flows in a shallow marine environment can be of different origin and may have an internal or external control. Mass transport flows are an example of episodic flows, for example turbidites. These types of mass transport flows are not known in the study area, they are however common in the North Sea Basin (Van der Molen, 2004). An external component is the influence of storms on sediments, this was already recognized by Zijlstra (1995). Lasseur et al. (2007) also recognized this in lower Cenomanian chalk in the Paris Basin (France). The presence of fossil-grit layers hence shows that the paleo-water depth was near or above storm wave base. Since the fossil-grit layers do not show any reworking by permanent flows, it is suggested that the shell concentrations were deposited below the fair-weather wave base (Reineck et al., 1973; Lasseur et al., 2007). A relative increase in grain size of the fossil clasts indicates relative higher energy conditions, which may be caused by a shallowing of the depositional environment or an increase in storm intensity (Zijlstra, 1995). The base of fossil-grit layers is an erosional surface. According to Myrow et al. (1991) erosional surfaces evolve from flat to undulating and eventually disappear with increasing depth and hence decreasing flow regime (Lasseur et al., 2007). Most erosional surfaces in the study area have an undulating top surface.

Lasseur et al. (2007) proposed a waterdepth model for the lower Cenomanian Chalk of the Paris Basin. The model is based on storm-induced shell concentrations in two components of successive metre-thick cycles: a depositional member and a top hiatal surface. All elements are similar to the Late Campanian and Maastrichtian biodetrital limestone in South Limburg, as well as the cyclicity of a depositional member and a top hiatal surface.

Seven depositional facies are defined, mainly based on the type of shell concentrations in the depositional members, as well as hiatal surfaces (Lasseur et al., 2007). The shell concentrations can be grouped into different units: SC 1 to SC 6. From SC 1 to SC 6, the mean grain size decreases, the fragmentation decreases, the packing decreases (from dense to dispersed) and the bioturbation increases (resulting in fading of clear bedding). The depositional facies range from coarse-grained highly bioclastic chalk (DF 1 and DF 2), through fine-grained moderately bioclastic chalk (DF 3 to DF 5) and to silt-grade ‘’muddy’’ chalk (DF 6 and DF 7). The shell content decreases with decreasing grain size (Lasseur et al., 2007).

For the hiatal surfaces eight different units are assigned, based on the firmness of the substratum (Lasseur et al., 2007). This ranges from softgrounds (SG 1 and SG 2) to firmgrounds (FG 1 and FG 2) up to hardgrounds (HG 1 to HG 4), which in turn are subdivided into subclasses. A field criteria for hiatal surfaces is the degree of compaction of burrows (mainly *Thalassinoides*), which is influenced by the degree of dewatering of the sediments. Highly deformed burrows, formed in not totally dewatered sediments, are indicative for softgrounds. Deformed burrows, formed in incomplete dewatered sediments, are indicative for firmgrounds. Undeformed burrows, formed in dewatered sediments, are indicative for hardgrounds (Lasseur et al., 2007).

Lasseur et al. (2007) positioned the storm wave base at the transition of SC 5 to SC 6, since these units do not show any traces of current activity (mainly mudstone facies and no shell concentrations). So SC 1 to SC 5 were deposited in the shoreface, while SC 6 was deposited in the offshore. The shell concentrations in the lower shoreface should show changes in the texture and grain-size as well as the nature of erosional surfaces, since the velocity of storm-related flows decreases with increasing water depth. This implies that the shell concentrations deposited during highest hydrodynamic energy conditions (SC 1) were deposited in the most shallow upper shoreface zone. As the shell concentrations decrease in mean grain size, fragmentation and packing, the depositional environment can be interpreted as a relative deepening. So from SC 1 to SC 5 the paleo-water depth is gradually increasing into the lower shoreface (Lasseur et al., 2007).

When using the water depth model of Lausseur et al. (2007), the Zevenwegen Member should be deposited in the offshore zone (table 5.3.1). The Beutenaken and Vijlen Member are deposited on the transition from offshore to shoreface, while the Vijlen Facies B was deposited in a relative shallower environment than Vijlen Facies A. During the Lixhe and Lanaye Member the position in the upper offshore is moving towards shallower depths. From the Maastricht Formation upwards, the paleo-water depth keeps decreasing from lower shoreface to upper shoreface.

Member	Shell concentr. unit	Hiatal surfaces unit	Depositional Facies
Zevenwegen	SC 6	none	DF 7
Beutenaken	SC 6 – SC 5	None	DF 6
Vijlen Facies A	SC 6	None	DF 6
Vijlen Facies B	SC 5	None	DF 6
Lixhe	SC 5	None	DF 6
Lanaye	SC4 and SCa	SG 2	DF 4
Maastricht Facies	SC 3 – SC 2 and SCa	HG 3 – HG 1 (Emael)	DF 3
Kunrade Facies	SC 3 – SC 2	HG 2 – HG 1	DF 2
Nekum	SC 1	HG 2 – HG 1	DF 2
Meerssen	SC 1	HG 1	DF 1

Table 5.3.1. The Members (Felder, 1975) of the study area subdivided by the water depth model of Lasseur et al. (2007).

This model corresponds with interpretations of the depositional environment by other authors for most members. However, the sediments of the Nekum and Meerssen Member are thought to be deposited under continuous influence of waves, which indicates deposition in the shoreface and foreshore. This depositional environment is not incorporated into the model of Lasseur et al. (2007). There is another discrepancy between the depositional environments in the Paris Basin and the study area. The Paris Basin is a basin subject to thermal subsidence (Lasseur et al., 2007), which results in a continuous subsidence with relative high subsidence rates. The study area is positioned on top of the London-Brabant Massif and sedimentation rates are relative low. So the Chalk in the Paris Basin was deposited in slightly deeper depositional environments than the study area and hence this model is only partly correct.

5.4 Paleoenvironmental reconstruction

Cretaceous sedimentation starts with deposition of fluvial and shallow marine clastic sediments in the Santonian (Aken Formation). Before that time an inverted Roer Valley Graben as well as an uplifted London-Brabant Massif resulted in non-deposition (Ziegler, 1990). The sediments of the Aken Formation were deposited during a transgression (Vandenberghe et al., 2004). The depositional environment changed from a fluvial depositional environment in the east of the study area to a shallow marine depositional environment towards the basin (Fig. 5.4.1). Also in the southwest of the study area, the London-Brabant Massif was not yet flooded and partly formed a terrestrial environment. The water depth during deposition can be interpreted as 20 - 40 m according to Felder et al. (2000). This is confirmed by the presence of cross bedding, lamination, channel incisions and tidal bundles.

During the Campanian, the siliciclastic sediments of the Vaals Formation were deposited in a shallow marine environment. The base of the sediments show a coarse-grained gravel bed of varying thickness, which indicates high energy depositions during deposition as well as erosion. This indicates that the Vaals Formation is deposited unconformable over the Aken Formation. Albers (1974) made a distinction into two different facies which show a decrease in energy conditions during deposition from east to west. This can be interpreted as a relative deepening of the depositional environment towards the west, which indicates relaxation of the London-Brabant Massif. The increase in $CaCO_3$ content indicates that conditions were increasingly favourable for carbonate production. The water depth during deposition can be interpreted as 20 to 50 m, based on the absence of sedimentary structures indicating flowing water on a regular basis. Incision of channels, bioturbation and lamination are common, indicating the presence of episodic flows.

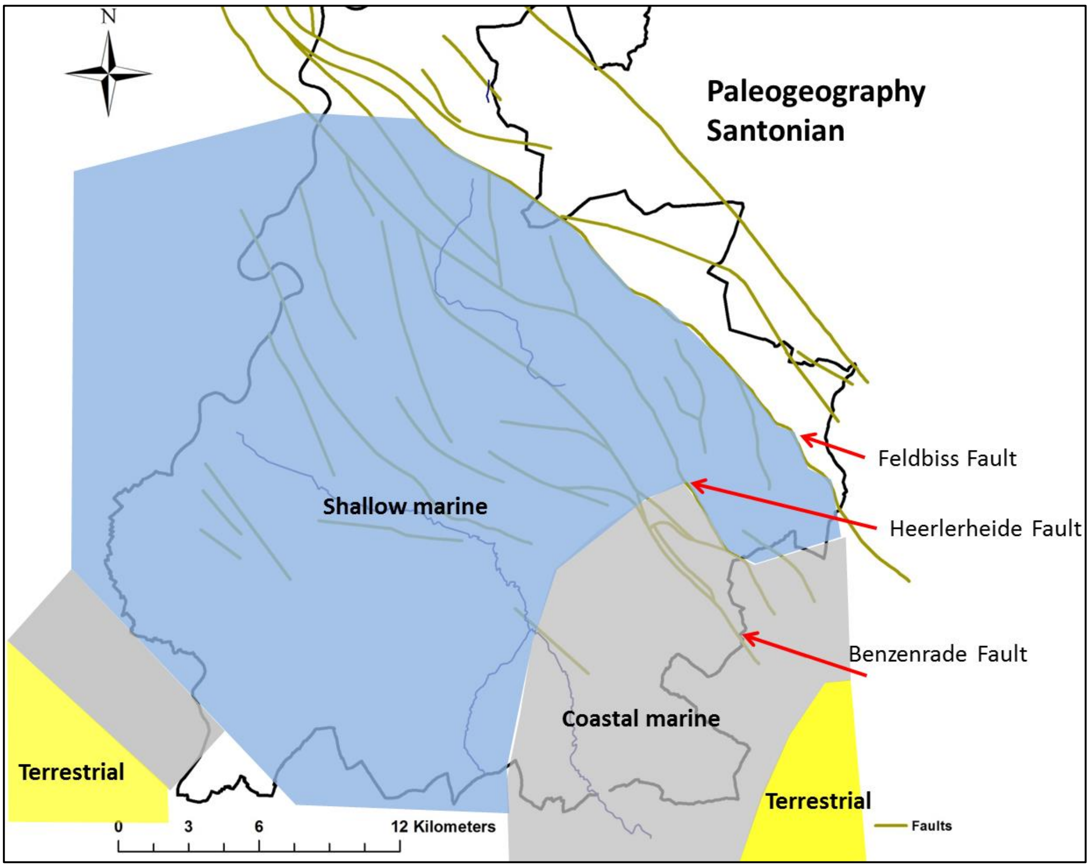

Fig. 5.4.1. Paleogeography during the Santonian, colours show different depositional environments.

Transgression continued during deposition of the Zevenwegen and Beutenaken Member. In a large part of the study area a biodetrital chalk, mainly composed out of coccoliths, was deposited below storm wave base (Fig. 5.4.2). Note that the term deep marine is relative (Fig. 5.4.2), these sediments probably formed at water depths below 50 m since no traces of episodic flows are observed. The gradual increase in $CaCO_3$ content from the northeast to southwest (up to 97%) can be interpreted as a relative distal depositional environment in the southwest, as only a small amount of fine-grained siliciclastic sediments were transported.

The base of both members contains a relatively large amount of fine-grained glauconite that decreases gradually towards the top. According to Felder et al. (2000) this glauconite was not formed in-situ, but eroded and transported from elsewhere. The observed gradual decrease would then indicate changes in denudation rates. This is likely caused by a decrease in uplift rates of the hinterland, which was formed by the inverted Roer Valley Graben during the Late Campanian. Another possibility is that the gradual decrease is caused by a decrease in distance to the source area. As the depositional environment is more proximal, detrital grains as glauconite are transported and fall out of suspension as the hydrodynamic energy conditions are gradually decreasing. The grain size is of importance since fine-grained particles are further transported than coarse-grained particles. This would indicate that large sea level variations operated on relative short time scales, which is well possible if relative sea level change is dominantly influenced by tectonics. Probably it has been a combination of relative sea level and the amount of erosion and transport of clastic grains.

Siliciclastic sediments with coarser-grained particles that show an increase in cementation of the sediments towards the top, are grouped as the Facies Zandig Krijt van Benzenrade (Fig. 5.4.2). This facies forms the more proximal equivalent of the Zevenwegen and Beutenaken Facies. These sediments are dominantly composed of clastic sand with glauconite. Recall that no glauconite is present in the Zevenwegen Facies, except the basal layer, while the Zandig Krijt van Benzenrade Facies does contain glauconite. This confirms the previous hypothesis.

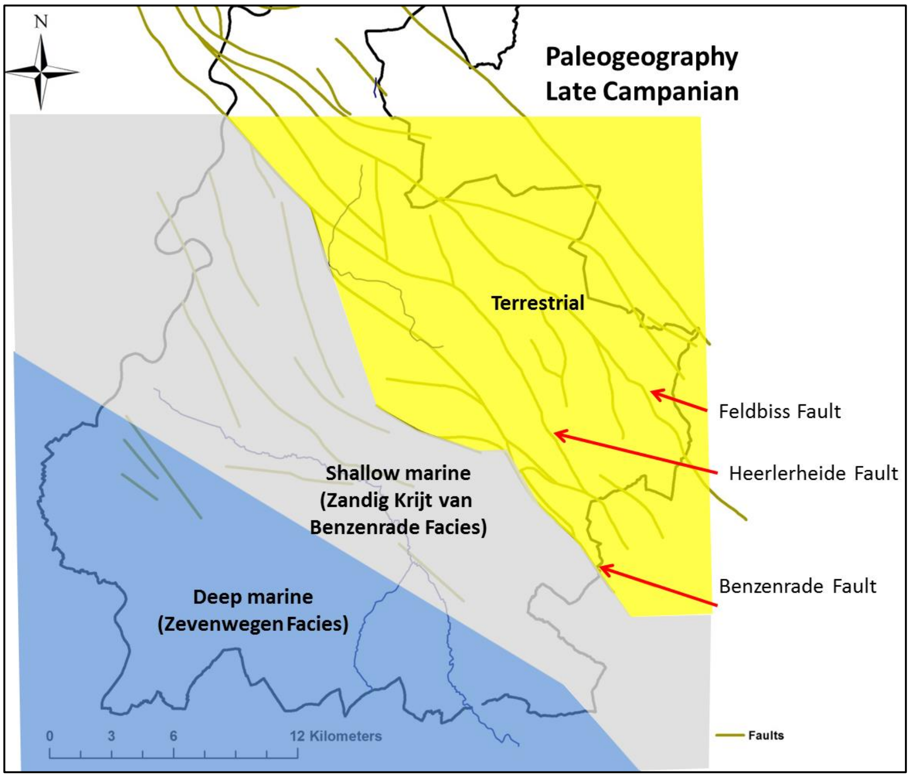

Fig. 5.4.2. Paleogeography during the Late Campanian, colours show different depositional environments.

Before deposition of the Vijlen Member, an event occurred that led to almost complete erosion of the Beutenaken Member, in a few locations the lower part of the Beutenaken Member is still present. In two zones of 1 to 2 km wide and multiple kilometres long erosion also removed older Cretaceous sediments. The cause of this will be discusses in another paragraph. During onset of sedimentation of the Vijlen Member, marly sediments were deposited formed during a transgression. Vijlen Facies B represents the lower part and consists of marls and calcisiltites with a large clay content, formed within a shallow marine depositional environment. Vijlen Facies A resembles a deeper marine depositional environment where calcisiltites with flint nodules are dominant. The $CaCO_3$ content also increases from 60 to 90% from Vijlen Facies B to Vijlen Facies A indicating a shift towards a more distal deposition environment. The northeast part of the study area contains sediments classified as Vijlen Facies B, while sediments in the west are classified as Vijlen Facies B in the lower part and Vijlen Facies A in the upper part. This suggests that proximal shore conditions were found in the northwest (inverted Roer Valley Graben), which is confirmed by an increase in coarse-grained siliciclastic fragments in fossil-grit layers and siliciclastic intervals, towards the Roer Valley Graben. Transgression during deposition of the Vijlen Member is shown by the gradual increase in $CaCO_3$ content, the shift of mollusc-dominated bioclasts to echinoderm-dominated bioclasts (Felder et al., 1989) as well as a vertical decrease in siliciclastic intervals and the grain size of clasts.

Transgression continued during deposition of the Lixhe Member as suggested by the relatively small grain size of bioclasts, the gradual increase in thickness of flint nodules and the gradual increase in $CaCO_3$. In the Lanaye Member the grain size of biodetrital clasts increases and hummocky cross stratification is occasionally observed. This suggest a shallowing of the depositional environment to depths around storm wave base. The flint nodules remain relatively large and the $CaCO_3$ reaches a maximum of 98%. This is interpreted as a distal depositional environment with conditions very favourable for carbonate production shown by the high $CaCO_3$. A more proximal facies is found in the Valkenburg Member around Valkenburg and in the Kunrade Facies, as is shown by correlations based on biostratigraphy (Felder et al., 1985 and 2001).

During deposition of the sediments of both the Kunrade and Maastricht Facies the depositional environment showed great differences (Fig. 5.4.3). The general trend is a shallowing of the depositional environment as shown by the gradual increase in grain size of the bioclasts. During deposition of the base of the Kunrade Facies, a calcarenite with glauconite and alternating cemented and less cemented beds formed. During the same time a calcarenite with a lower glauconite content, was deposited in the area of Valkenburg (lower part Valkenburg Member). It shows irregular layers and lenses of cemented beds. The resemblance between both locations is that the sediments contain a relatively large glauconite concentration at the base that is gradually decreasing upwards. Further towards the west calcarenites were deposited, thick flint-nodule layers later formed during diagenesis . In a later period, deposition of calcarenites continued in the Kunrade area, while in the area of Valkenburg calcarenites formed, during diagenesis randomly distributed flint nodules formed (upper part Valkenburg Member, Gronsveld Member and Schiepersberg Member?). In the area of Maastricht, calcarenites formed with distinct and continuous flint nodule layers that can be tracked laterally towards the south and west (Valkenburg Member, Gronsveld Member and Schiepersberg Member). The upper part of the Kunrade Facies unit is similar to the sediments below, in the Valkenburg and Maastricht area calcarenites were deposited on top of a distinct fossil-grit layer (Horizon of Romontbos) that show specific pipe and plate-shaped flint nodules.

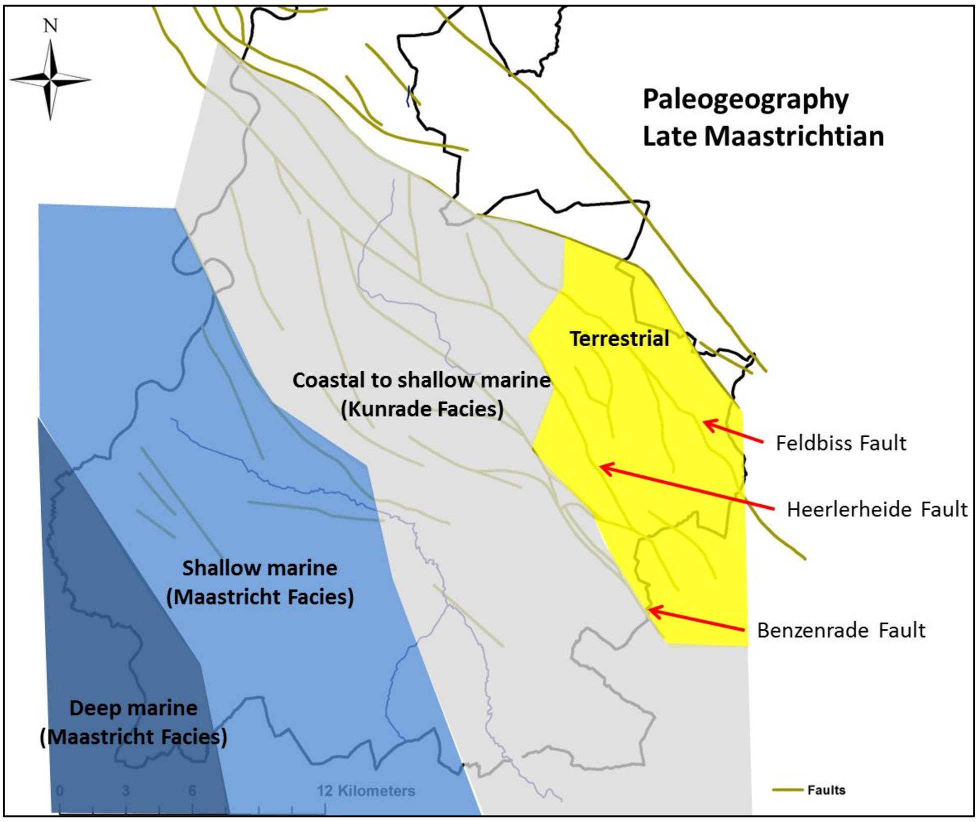

Fig. 5.4.3. Paleogeography during the Late Maastrichtian, colours show different depositional environments.

During deposition of sediments of the Nekum and Meerssen Members, shallowing of the depositional environment continued as shown by the relative increase in grain size of biodetrital clasts. Flint is only present in the lower part of the Nekum Member. The highest amount of flint nodules is found in Nekum Facies A which resembles a relative deeper facies than Nekum Facies B. This deepening trend from Nekum Facies A to B is also confirmed by the degree and amount of cementation, which is largest in relative shallow marine environments subject to regular wave activity (Inden et al., 1983). The presence of regular wave activity can also be deduced from the presence of many thin fossil-grit layers that resemble storm deposits (Zijlstra, 1995). Relatively more fossil-grit layers can be observed in Nekum Facies B than Nekum Facies A, again confirming the relatively shallow depositional environment of Nekum Facies B. The distribution of both facies suggests that the Roer Valley Graben remained a zone with relatively shallow marine conditions and a source of clastic material as shown by the decrease in $CaCO_3$ content towards the northeast. The sediments of the Meerssen Member show that the depositional environment was more shallow marine than for the sediments of the Nekum Member. This is evidenced by the common appearance of hardgrounds and corresponding fossil-grit layers. Since there is a positive relation between the degree of cementation and the duration of non-deposition, which is in turn positively related to hydrodynamic energy conditions, it can be concluded that the sediments of the Meerssen Member were deposited in a shallower depositional environment than previously described members.

5.5 Facies Analysis

The different facies of the Gulpen and Maastricht Formations can be group into the following five facies associations.

Fine-grained carbonate facies
Sediments consist of fine-grained carbonate mudstones to wackestones with a varying $CaCO_3$ content that ranges from 75% at the base to 97% at the top. In the upper part small flint nodules may be present, dominantly in the western part of the study area. The sediments of the Zevenwegen, Beutenaken and Vijlen Members (Felder, 1975) can be classified as this facies. Each lithostratigraphic unit shows a several metres thick basal layer that shows a high glauconite concentration which gradually decreases upwards. The sediments of the Vijlen Member contain a relative large amount of glauconite which indicates increased erosion rates during deposition. In the top of each lithostratigraphic unit a hardground has formed which indicates a hiatus in between the lithostratigraphic units.

The depositional environment of the sediments of this facies unit can be interpreted as a low-energy and distal environment, as shown by the lack of sedimentary structures and relative low siliciclastic content. The paleo-water depth can be interpreted as below the storm wave base since no sedimentary structures created by waves or flows can be observed.

Calcarenite with distinct flint nodule layers facies
Sediments consist of calcarenites. Multiple flint nodule layers that are continuous can be distinguished and increase in thickness upwards. The $CaCO_3$ content ranges from 80% at the base to 98% at the top. The sediments of the Lixhe, Lanaye, Valkenburg, Gronsveld, Schiepersberg and Emael Members (Felder, 1975) can be classified as this facies. It is only found in the southwest of the study area around the valley of the river Maas and west of it.

The depositional environment of the sediments of this facies unit can be interpreted as a low-energy to high-energy environment. The relative high $CaCO_3$ points to a relative distal depositional environment. The paleo-water depth can be interpreted as around the storm wave base since sedimentary structures created by waves or flows can be observed. Hummocky cross stratification, cross bedding and lamination is occasionally visible. Fossil-grit layers at the base of each lithostratigraphic member can be interpreted as storm deposits.

Calcarenite with fossil-grit layers and randomly distributed flint nodules facies
The sediments consists of calcarenites with many flint nodules, all randomly distributed. Coarse-grained fossil-grit layers are common and have a thickness of multiple centimetres to decimetres. The sediments show a faint alternation of hard and soft beds, of which hard beds have a higher $CaCO_3$ content. The $CaCO_3$ content ranges from 70% to 90%. The sediments of the Lixhe, Lanaye, Valkenburg, Gronsveld, Schiepersberg and Emael Members (Felder, 1975) can be classified as this facies. It is only found in the middle part (Cadier & Keer and Valkenburg) and in the east (Gulpen) of the study area.

The depositional environment of the sediments of this facies unit can be interpreted as a relative high-energy environment. The paleo-water depth can be interpreted as above the storm wave base since sedimentary structures created by waves or flows are common. Cross bedding and lamination is occasionally visible. Coarse-grained fossil-grit layers at the base of each lithostratigraphic member and throughout the sequence can be interpreted as storm deposits.

Calcarenite with cemented beds facies
The sediments consist of calcarenites with an alternation of slightly undulating hard and soft beds, occasionally flint nodules are randomly present. Coarse clastic grains and fossil-grit layers are common as well as glauconite. The $CaCO_3$ content ranges from 69% in soft beds to 98% in hard beds.

The sediments of the lower part of the Maastricht Formation in the east, also known as the Kunrade Facies can be classified as this facies. It is only found in the east of the study area, near the Roer Valley Graben.

The depositional environment of the sediments of this facies unit can be interpreted as a high-energy environment. The alternating $CaCO_3$ content points to a cyclic alternation in supply of clastic grains, which may be precession-related. The paleo-water depth can be interpreted as above the storm wave base since sedimentary structures created by waves or flows can be observed. Cross bedding and lamination is occasionally visible. The undulating beds, as well as the sedimentary structures indicate that deposition occurred near the fair weather wave base.

Calcarenite with hardgrounds and fossil-grit layers facies
The sediments consists of calcarenites with a $CaCO_3$ content that ranges from 96% to 98%. In the lower part small flint nodules may be present, dominantly in the western part of the study area. Towards the top hardgrounds become more common. Fossil-grit layers are relatively thick, up to tens of centimetres, and coarse-grained. The sediments of the Nekum and Meerssen Members (Felder, 1975) can be classified as this facies. Each lithostratigraphic unit shows a several decimetre thick coarse-grained basal fossil-grit layer.

The depositional environment of the sediments of this facies can be interpreted as a high-energy environment, as shown by the presence of many fossil-grit layers that show lamination or cross bedding. The paleo-water depth can be interpreted as around the fair-weather wave base lamination and cross bedding can be observed.

5.6 Relation (inversion) tectonics and sedimentation

There is a direct relation between tectonics and sedimentation in the study area, this is enhanced by the relative shallow depositional environment which is easily influenced by relative sea level fluctuations.

Sediments of Campanian age are only found in the south (Zevenwegen and Beutenaken Members) and east (Zandig Krijt van Benzenrade unit) of the study area. North of Maastricht and around Valkenburg these sediments are missing as sediments of Maastrichtian age (Vijlen Member) are found directly on top of Early Campanian sediments (Vaals Formation). This may points to non-deposition or erosion after deposition. Non-deposition is unlikely since the Zevenwegen and Beutenaken Members are interpreted as sediments deposited in a relatively deep depositional environment which would have resulted in sedimentation in most areas. Hence, erosion is more likely. The sediments of the Vijlen Member also point to a period of erosion before deposition of the Vijlen Member. Two deeply incised elongated zones, parallel to the Roer Valley Graben, were found filled with a thick layer (50 to 60 m) of sediments classified as the Vijlen Member. The Vijlen Member of Early Maastrichtian age is found on top of Carboniferous strata, while sediments of Santonian and Campanian age are missing. This indicates that erosion was more pronounced in the northwest and in the erosional zones. The upper part of the Beutenaken Member is absent while the lower part is only present in two areas. This points to a relative sea level drop resulting in erosion. The relative sea level drop is caused by changes in tectonic regime, presumably increased rates of uplift due to an increase in compressional forces. The rates of uplift varied as is shown by the thickness distribution of the different units of Campanian age. Rates of uplift must have been large in the northwest (north of Maastricht and Valkenburg area) since Late Campanian sediments are missing. Towards the south and east uplift must have been less important as the oldest Late Campanian sediments are present

(Zevenwegen Member). In the elongated erosional zones, uplift rates were highest as no Late Cretaceous sediments are preserved. Overall, the study area has been subject to increased rates of uplift at the transition from Campanian to Maastrichtian. This also caused the formation of hiatuses in between the different lithostratigraphic members (Fig. 5.6.1). This figure clearly shows that only a limited amount of time is represented in the study area.

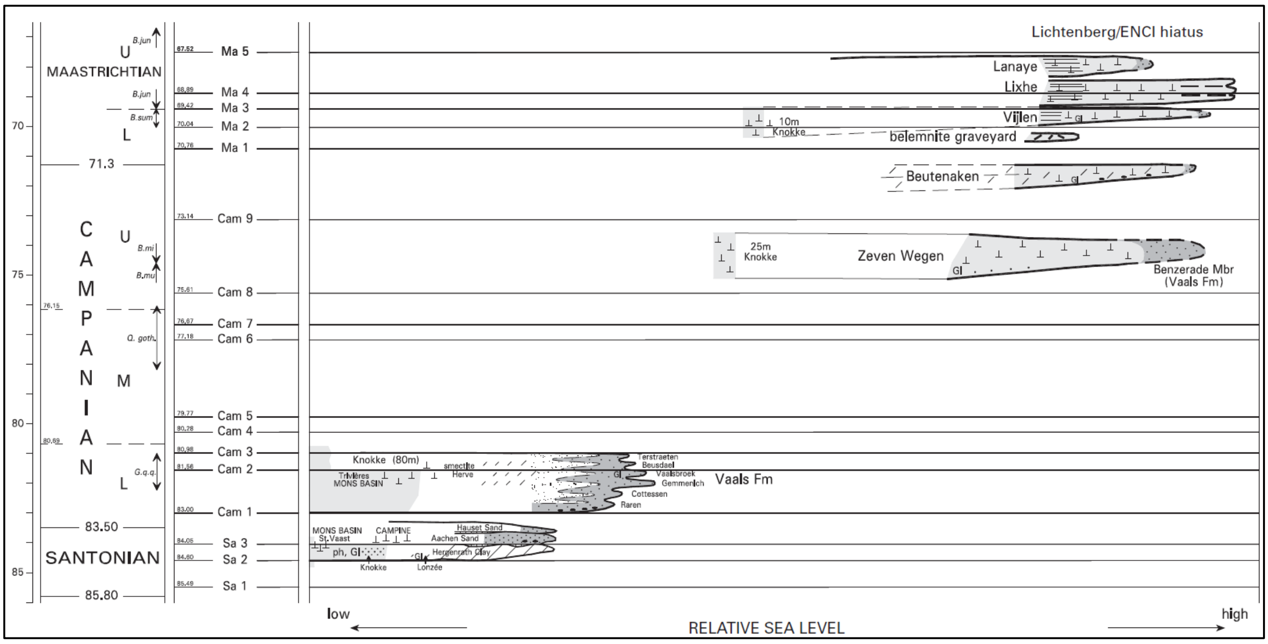

Fig. 5.6.1. Aken and Vaals Formations and the individual members of the Gulpen Formation in a chronostratigraphic chart. Note that only a limited period of time is represented by sediments in the study area. (From: Vandenberghe et al., 2004)

The different rates of uplift can be explained if a horst and graben structure is accepted on the southwest shoulder of the inverted Roer Valley Graben. The result of uplift and associated erosion is visualized in the cross-sections (Appendix). Profile 2 (Appendix) is drawn through the area of Valkenburg where the Campanian sediments are missing, and clearly shows a large variation in thicknesses of different units along faults. The strata in a specific area bounded by faults is described as a block in this research. As profile 2 shows each block shows different thicknesses per unit, as well as different trends in thickening. This shows that the inversion tectonics of the Roer Valley Graben are quite complex as each block responded differently. It can be described as a horst and graben structure which caused exposure and erosion in one location while sedimentation prevailed in another location. The variation in thickness varies per unit and per block, which indicates that tectonic forces varied over short time spans (Fig. 5.6.2).

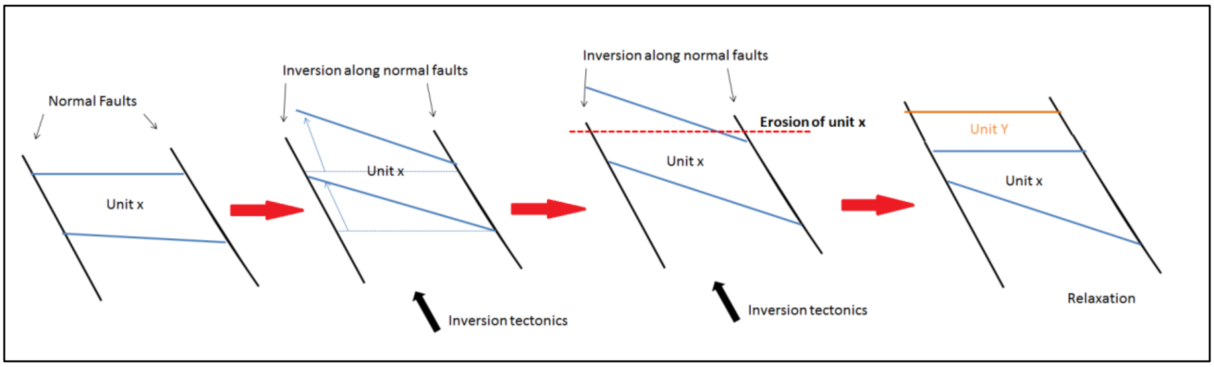

Fig. 5.6.2. Sketch of strata in between normal faults being inverted and truncated.

Erosion of sediments must have resulted in transport and sedimentation elsewhere. In the North Sea region, Late Cretaceous sediments are occasionally redeposited by mass flow deposits (Wray et al., 2006). These can be turbidites (Jones et al., 1992), debris flows (Gale, 1980) and slumps (Wray et al., 2006). No evidence for these mass flows has been found in the study area. This is logical to some extent since deposition occurred towards the basin centre which is not in the study area. The erosional zones may have been the result of differential uplift of blocks along the rim of the Roer Valley Graben. However, no faults have been interpreted in or near these zones, there is however some evidence of fault activity. Faults do not have to be visible necessarily, flexure of the overlying sediments caused by blind thrusts in the subsurface is also possible. After uplift, truncation of the sediments occurred followed by subsidence of the strata Unfortunately no seismic data is available. The parallel orientation of the erosional zones with the Roer Valley Graben is in agreement with a horst and graben structure caused by inversion tectonics.

After the period of increased uplift during the transition from Campanian to Maastrichtian, tectonic movement decreased and thicknesses of units show less variance. This is shown by the absence of areas where erosion occurred.

6. Conclusion

The aim of this study is to discuss the sedimentological significance of the Late Cretaceous sediments in the Dutch part of the Campine Basin. To do so, a facies analysis is presented which shows new insights about the depositional environment and the tectonic setting of the structural elements in the study area.

The siliciclastic sediments of the Aken and Vaals Formation were mainly deposited before Campanian inversion of the Roer Valley Graben. The depocenter was directed towards the eastern part of the Roer Valley Graben. During deposition of the calciclastic Gulpen and Maastricht Formation the depocenter switched towards the west. This was caused by an inversion of different fault blocks on the southeast shoulder of the Roer Valley Graben, as well as increased subsidence rates in the Campine Basin towards the west. The uplift associated with the inversion resulted in several erosional phases, mainly concentrated in the north part of the study area. Along the Heerlerheide and Benzenrade Faults most movement occurred, as is shown by great variations in thicknesses of different units along both sides of these faults. It also appears that uplift, and associated erosion, occurred in areas where no faults have been interpreted. Presumably sediments have been eroded due to flexural movement or movement along blind thrusts. Future research might solve this.

The biodetrital sediments of the Gulpen and Maastricht Formations increase in siliciclastic content towards the Roer Valley Graben. This indicates that the Roer Valley Graben still formed a topographic high where erosion prevailed. The Maastrichtian sediments clearly show a gradual shallowing of the depositional environment.

Different facies associations have been distinguished within sediments of the Gulpen and Maastricht Formations. Almost all sediments are characterized by biodetrital clasts, which indicates reworking. This is the largest difference with chalk in basins of the North Sea where it largely forms a pelagic sediment. The most distal and deepest depositional environment is characterized by fine-grained chalk. In the upper offshore and lower shoreface, calcisiltites and calcarenites formed, during diagenesis distinct flint nodule layers formed. In the lower shoreface a calcarenite formed with fossil-grit layers, flint nodules formed during diagenesis but in less distinct layers. In the upper shoreface a calcarenite formed with coarse-grained fossil-grit layers, randomly distributed flint nodules formed during diagenesis. The foreshore is characterized by calcarenites with many fossil-grit layers and hardgrounds.

7. Recommendations for further research

The benthic foraminifera species of the different assemblage zones of Hofker (1966) have not been studied in detail. Since benthic foraminifera species are facies dependant, these can be used to interpret the depositional environment. When the relative abundance of specific species can be matched with a facies model than one can use the relative abundance of specific species to interpret the depositional environment. This can be a useful proxy in boreholes of the Roer Valley Graben where the sediments cannot be studied in detail. A correlation with other Cretaceous basins in northwest Europe would as well be interesting to see how different species differ in different areas.

A microscopic study of these sediments is missing, it could provide useful information about bioclasts, the type of cement, and siliciclastic grains. Thin sections were not available in this study but are of uttermost importance to study gradual changes in lithology. If several thin sections where obtained of coeval sediments in different locations, then a quantitative study could aid this facies analysis.

The Gulpen and Maastricht Formations are also found towards the southwest and the west of the study area (Belgium). However, these have not been studied and could contain useful information for the facies analysis.

Sedimentation of the Late Cretaceous sediments was influenced by tectonics of the Roer Valley Graben and the London-Brabant Massif. Several erosional features have been observed that cannot be explained with the existing fault model. Seismic research of the subsurface may supply additional information about the presence of faults and how these faults responded to changes in tectonic regime.

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9. Appendix

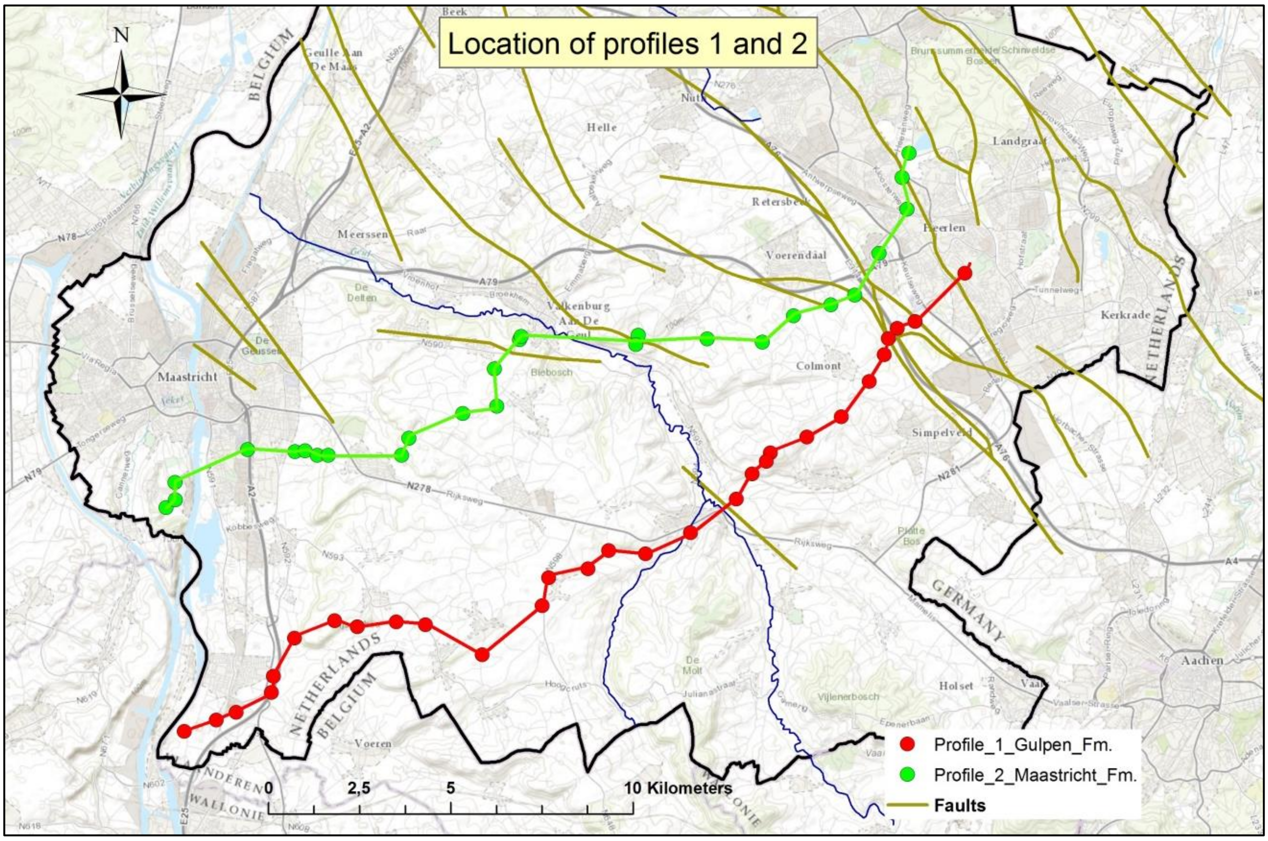
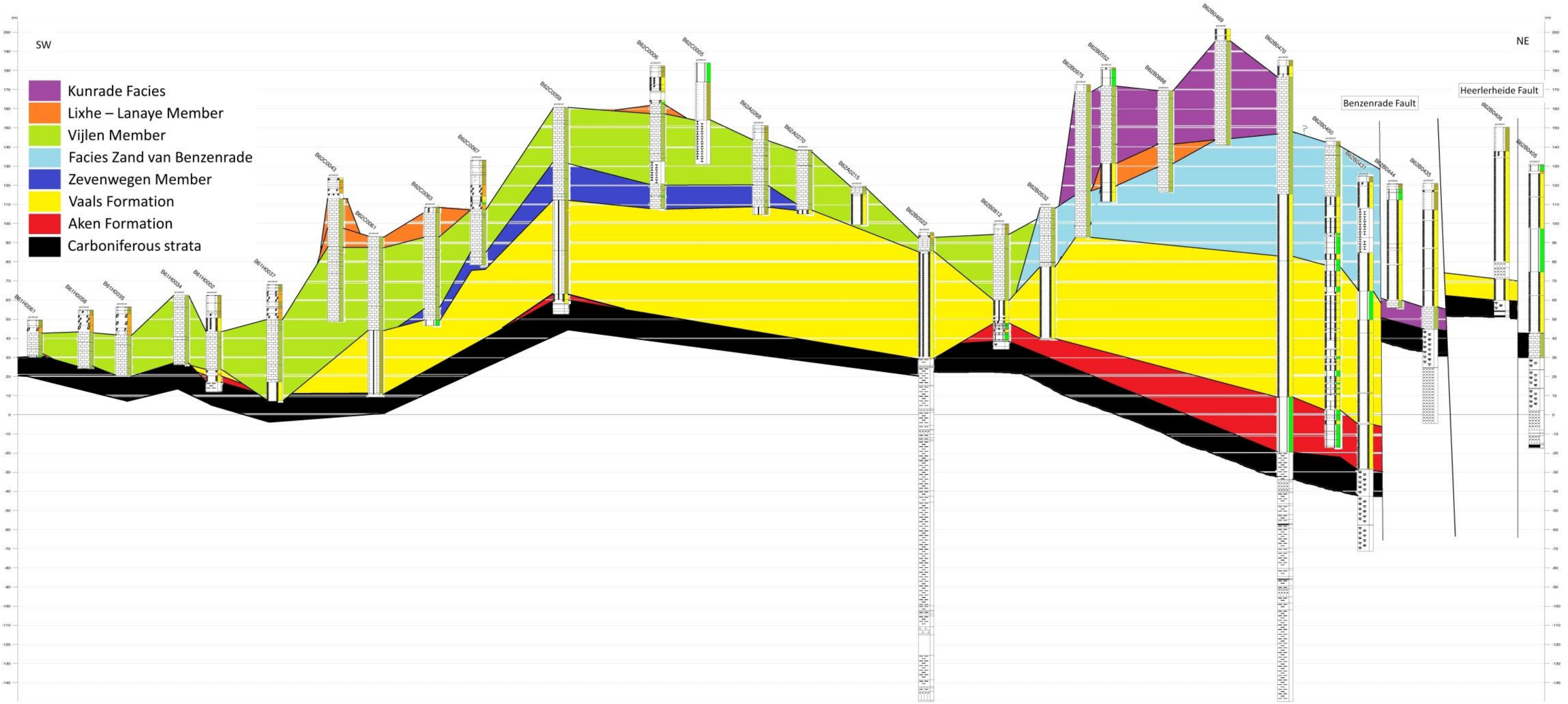
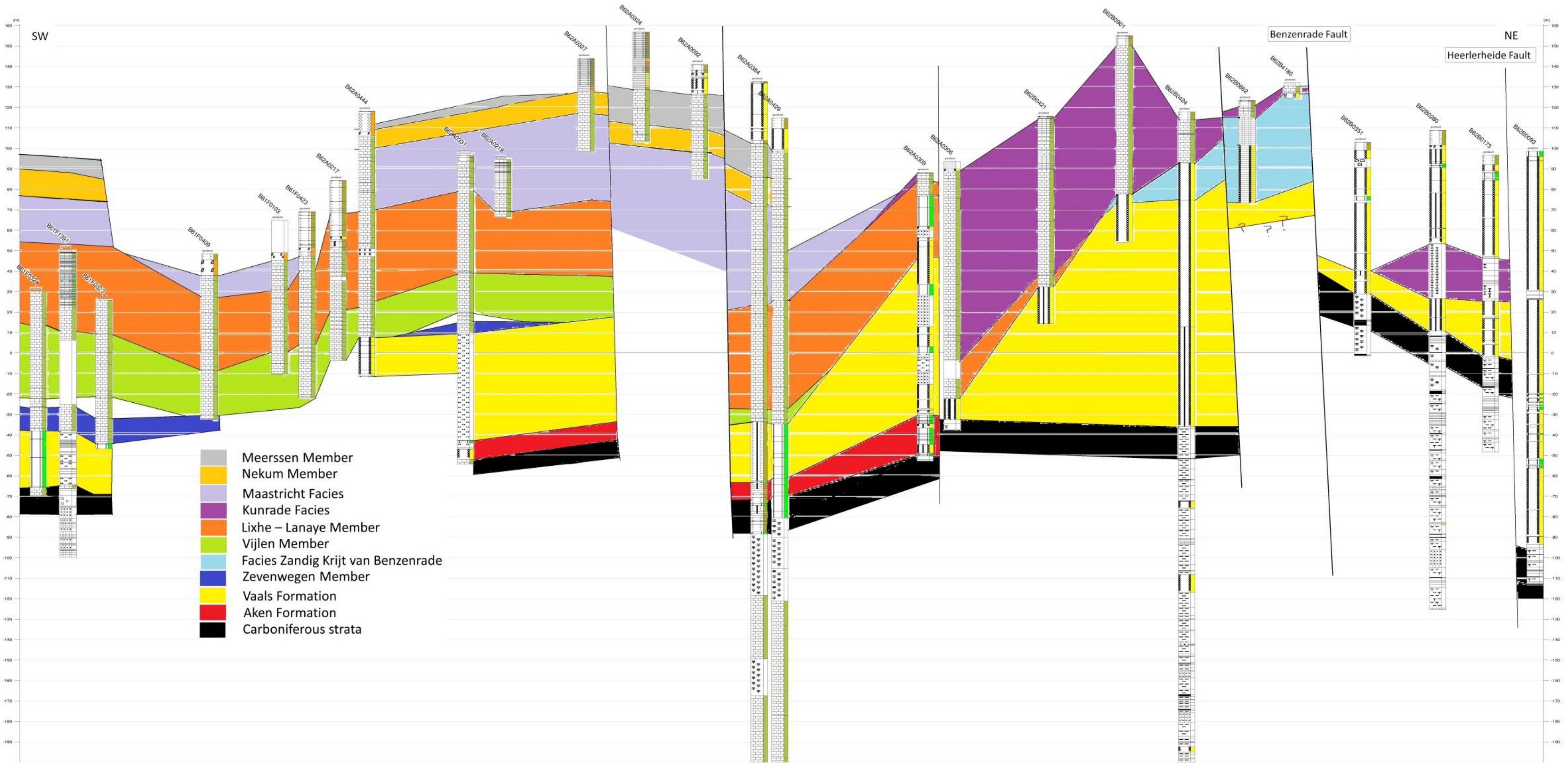


Fig 8.1. Location of profile 1 and 2.



Profile 1. Cross-section from the subsurface, from the line Eijsden, Gulpen and Heerlen. See fig. 8.1 for exact location of cross-section.



Profile 2. Cross-section from the subsurface from the line ENCI quarry (Maastricht), Valkenburg and Heerlen. See fig. 8.1 for exact location of cross-section.